



# A Natural Resource Condition Assessment for Sequoia and Kings Canyon National Parks

## *Appendix 7a – Water Quantity: Rain, Snow, and Temperature*

Natural Resource Report NPS/SEKI/ NRR—2013/665.7a



**ON THE COVER**

Giant Forest, Sequoia National Park  
Photography by: Brent Paull

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## Scope of Analysis

Precipitation, snowpack and air temperature, the primary physical determinants of climate, are the foundation for multiple focal resources in the Parks. Over thousands to millions of years they interact with the geologic template to determine soil formation and weathering. Over seasonal to decadal time scales they determine ecosystem health and the distribution of species. Temperature determines the elevation of rainfall versus snowfall, and thus the timing, partitioning, and seasonal magnitude of fluxes making up the hydrologic cycle in the Parks.

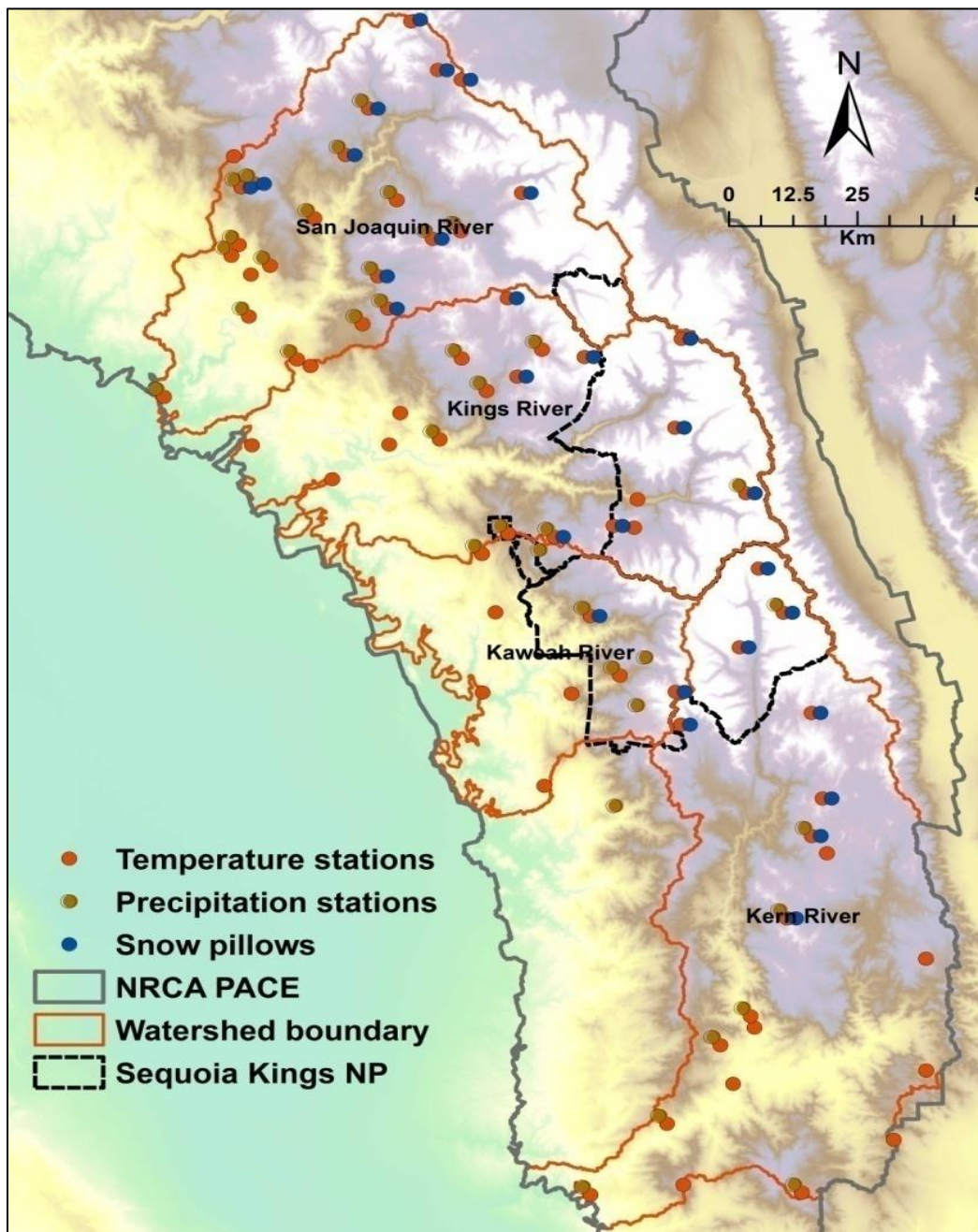
The current analysis included retrieving, assembling, documenting, and doing quality assurance/and quality control of measured temperature, snow, rainfall, and related paleoclimatic data for the four main river basins that SEKI lies within, the San Joaquin, Kings, Kern, and Kaweah River basins (Table 1, Figure 1). Part of the Tule River basin is also within SEKI, but it was not included in the analysis due to the small area that is within the boundary of the park (20 km<sup>2</sup>) and the similarities in topography to the Kaweah basin. The San Joaquin drains into the San Joaquin River of the in the Central Valley, while the Kings, Kaweah, and Kern are normally endorheic basins, though they do empty into the San Joaquin in exceptionally wet years with intense spring runoff. The Kings and Kaweah formerly emptied into Tulare Lake and the Kern formerly emptied into Buena Vista Lake at the southern end of the Central Valley. The total area of SEKI is 3502 km<sup>2</sup>. Since 91% of the elevations are above 1800 m, with a snow/rain transition zone of 1500-1799 m, seasonal snow is the fundamental driver of the hydrologic processes in SEKI.

**Table 1.** Basin areas and elevations

Basin	Above confluence		Within parks	
	Area, km <sup>2</sup>	Elevation, m	Area, km <sup>2</sup>	Elevation, m
<b>San Joaquin</b>	4418	92-4238	169	2447-4215
<b>Kings</b>	4789	110-4302	1641	1199-4302
<b>Kaweah</b>	2428	125-3814	907	414-3814
<b>Kern</b>	6142	154-4409	765	1929-4409
<b>Total</b>	17,777	92-4409	3482	414-4409

This chapter focuses on one subset of the water cycle within SEKI: snowpack, snowmelt, and precipitation, with the latter being the sum of rainfall and snowfall (Figure 2). These fluxes, plus snowpack and soil-water storage, determine the amount of water available for evapotranspiration (sum of evaporation and transpiration), and streamflow. Snowpack is a critical SEKI resource that provides seasonal storage of water for soil moisture, lakes and streams. It is particularly important given the seasonal cycle of precipitation and long summer-fall dry period in the southern Sierra Nevada. Soil moisture is also a critical SEKI resource, which also provides seasonal storage for sustaining lakes, streams and vegetation year round. The results presented here are based largely on data plus simple models, and set the stage for a further analysis of other water-cycle fluxes and reservoirs, e.g. evapotranspiration, streamflow, and soil moisture. Data to accurately estimate soil moisture and evapotranspiration (ET) across the landscape are generally not available, and streamflow is available at only a few locations in SEKI. Extending all of these quantities across the landscape will require a more-detailed hydrologic model, combined with

characterization of vegetation and physiographic features of the landscape. Thus a more-extensive analysis of the water cycle, while clearly needed to assess critical Park resources, is beyond the scope of the current analysis.

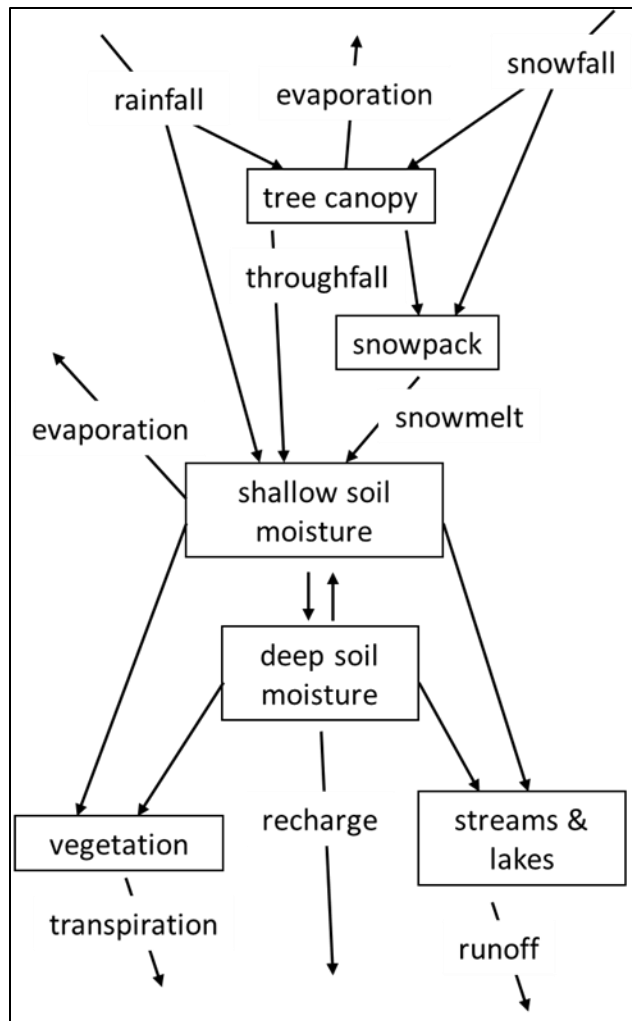


**Figure 1.** Study area showing temperature, snow pillow, and precipitation measurement sites in the San Joaquin, Kings, Kaweah and Kern River basins. Data from these sites were used in the current analysis.

Measurements by multiple entities resulted in the patterns and trends reported here. These measurements are also compared with the commonly used PRISM products for temperature and precipitation. PRISM (Parameter-elevation Regressions on Independent Slopes Model) is an interpolated product that uses some of the same point measurements of precipitation and



temperature analyzed here, plus data from the larger region, to produce continuous, digital grid estimates of monthly, yearly, and event-based climatic parameters (<http://www.prism.oregonstate.edu/>).



**Figure 2.** Main fluxes and reservoirs of interest in mountain water cycle in SEKI.

In the current analysis we chose three years to examine in more detail, representing wet (2005), average (2008) and dry (2007) years. The analysis focused on time series data for the period of record of the data. A brief analysis of available paleoclimate data is also included, to provide a longer-term context. Finally, an assessment of snowcover is provided by HUC 10 watershed within the parks.



## Critical questions

The central critical question addressed in this chapter concerns the extent and timing of snow-covered area, snowpack water content, and snowmelt. These quantities are used to define the SEKI snowpack resource. As precipitation in SEKI is a mix of rain and snow, an assessment is also done on total precipitation. Along with temperature, precipitation data are helpful in quantitatively assessing the snowpack resource.

Temperature is the main climate variable that is projected to rise significantly in the coming years and decades. Temperatures in California have risen about 0.6-1.0°C over the past century, comparable to the change in global average temperature (Figure 3a). Temperature is projected to rise another 1-4°C by mid-century and 2-8°C by 2100 (Figure 3b). Though the range of possible temperature scenarios covers several degrees, there is a strong consensus around these scenarios for northern California (e.g. Dettinger, 2005; Hayhoe, 2004; Cayan et al., 2008). Note that as a society, we are currently following the highest emissions scenarios on Figure 3b. In contrast, there is less clarity about changes in precipitation amount, though more scenarios suggest a slightly drier versus wetter mean by 2100 (e.g. Dettinger, 2005).

In relation to the precipitation quantities considered in this chapter, temperature aloft is the main climate variable in determining whether precipitation falls as snow or rain, ground-level temperature is central to determining the timing and rate of snowmelt, and the air temperature profile in the tree canopy (5-50 m) affects the rate of evapotranspiration in forests. Note that forests dominate the landcover in many parts of SEKI; and that ET in forests is much higher than for herbaceous vegetation. Thus understanding and predicting evapotranspiration by forest vegetation is central to understanding the effects of temperature and precipitation changes on the water cycle. Knowing how these quantities will change in the future is generally based on the assumption that one knows these quantities at present. Knowledge of precipitation and snowpack in the Sierra Nevada is weak owing to few measurements, especially measurements longer than a few years. Thus the first critical question concerns the magnitude of precipitation and snowpack across the Parks, given the complex topography and large elevation differences in the Southern Sierra Nevada. While climate projections of precipitation changes are mixed, and have greater uncertainty than those for temperature, the magnitude of changes from snow to rain with rising temperature is a large question that will affect Park management.

Spatial data on snow go back to 1999, with the launch of NASA's MODIS satellite. Thus the current analysis focuses more on that than on earlier periods, which also corresponds to a period of improved ground-based measurements in the Parks.

## Data sources and types used in analysis

Temperature, snow-pillow, and precipitation data were downloaded from the California Data Exchange Center (CDEC) (<http://cdec.water.ca.gov/>) and from the Western Regional Climate Center (WRCC) (<http://www.wrcc.dri.edu/>) for water years 2000-2009. Data were inspected for missing values, and checked for spikes and spurious points. Spikes and spurious data were eliminated and gaps filled by linear interpolation or correlation with nearby stations. If necessary the data sets were converted to hourly from 15-minute time series. The cleaned data

are archived at the Sierra Nevada San Joaquin Hydrologic Observatory (SNSJHO) Digital Library (<https://eng.ucmerced.edu/snsjho>) and archived in 3 levels: level 0 is the raw data downloaded directly from CDEC and WRCC; level 1 is the reformatted and quality controlled data; and level 2 is the gap-filled data.

## Temperature

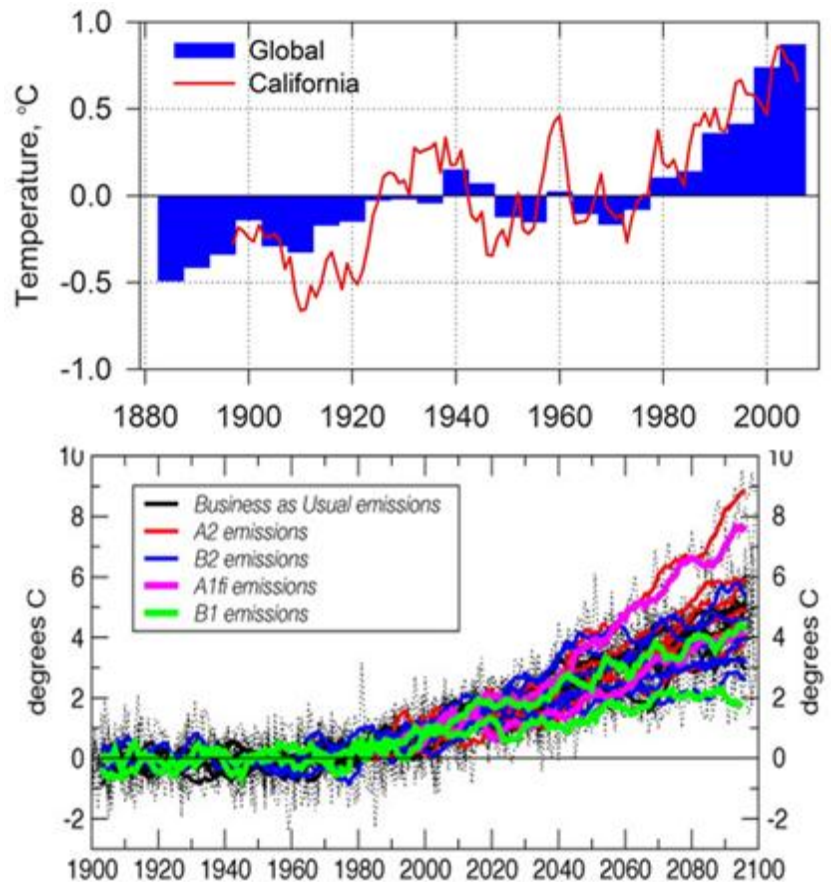
Temperature was measured at 71 locations in the basins by a variety of operators and with various levels of quality control with elevations ranging from 174 m to 3477 m. CDEC, which serves operational hydrology in California, archives temperature records for the San Joaquin, Kings, Kaweah, and Kern River basins reporting hourly, daily averages, minimum and maximum temperature (Figure 1). Most stations report hourly temperature for the period of record; and some report daily minimum, maximum, and average temperatures. Note that there is a lack of stations with consistent temperature data for the full 10-year period of WY 2000-2009.

Four stations in the Park have multi-decade records, though one station is inactive (Figure 4). Records are not sufficiently long to establish reliable trends, though an analysis to evaluate the increase in temperature over the past 4 decades could be carried out after cleaning of data.

Shorter-length temperature records from many of the stations also show clear annual cycles (Figures 5a-5d). Also apparent were some quality control issues that could not easily be cleaned in the data. For example, the Mammoth Pass data for 2007-present were adjusted by 15°C to match prior-year patterns; other anomalies that were not adjusted appear in 2006.

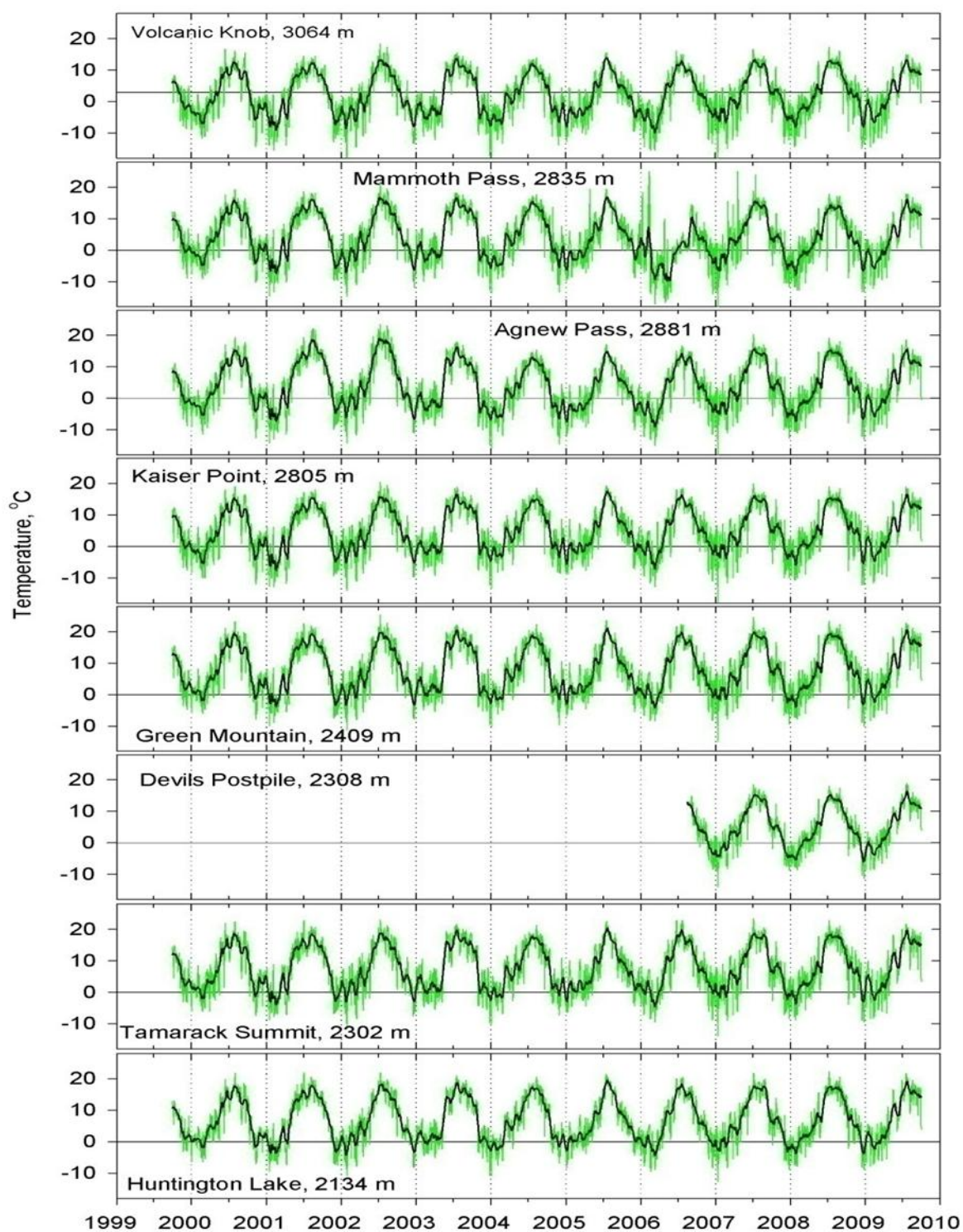
PRISM temperature (minimum and maximum) data for the 800 m normals (1971-2000) and the 4 km monthly (2000-2009) were downloaded from (<http://www.prism.oregonstate.edu/>).

Average PRISM temperatures across the 4 basins were developed from the PRISM minimum and maximum temperature gridded 800-m and 4-km products.

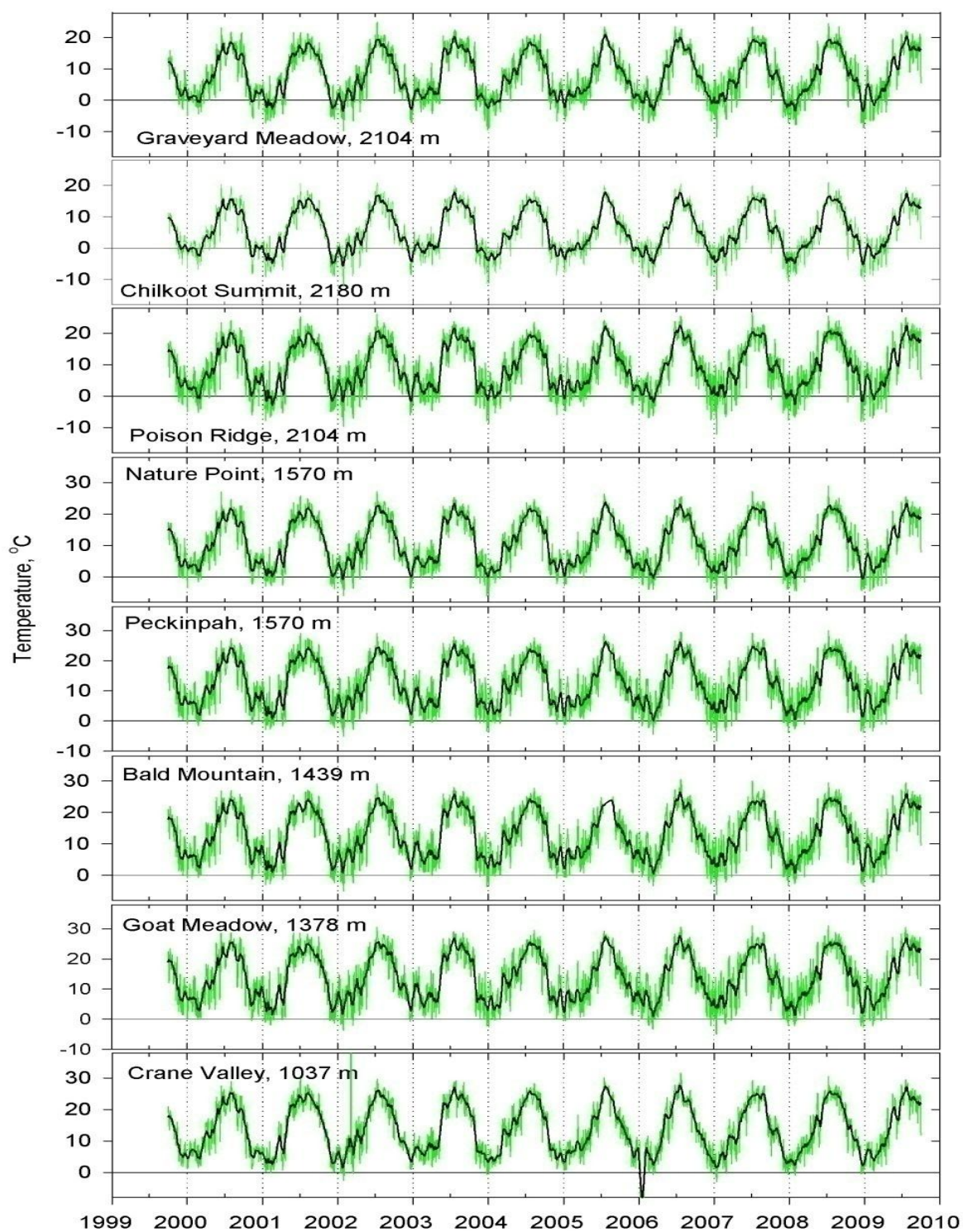


**Figure 3. a)** Global and California temperature history and **b)** projected temperature changes for California under various emissions scenarios (Dettinger, 2005).



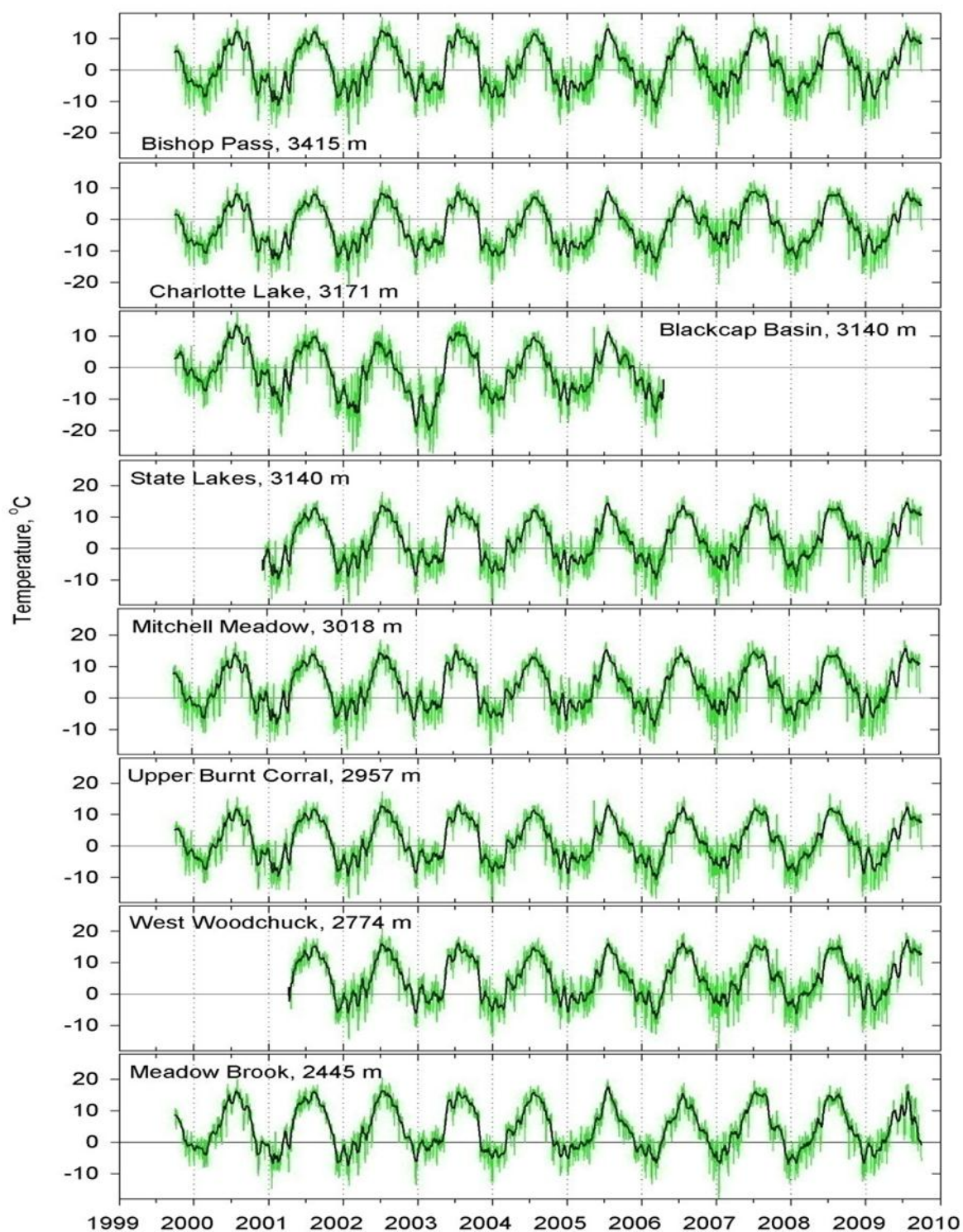


**Figure 5a.** Daily average temperature for 16 stations in the San Joaquin river basin. Lines are daily and a 20-day running mean.

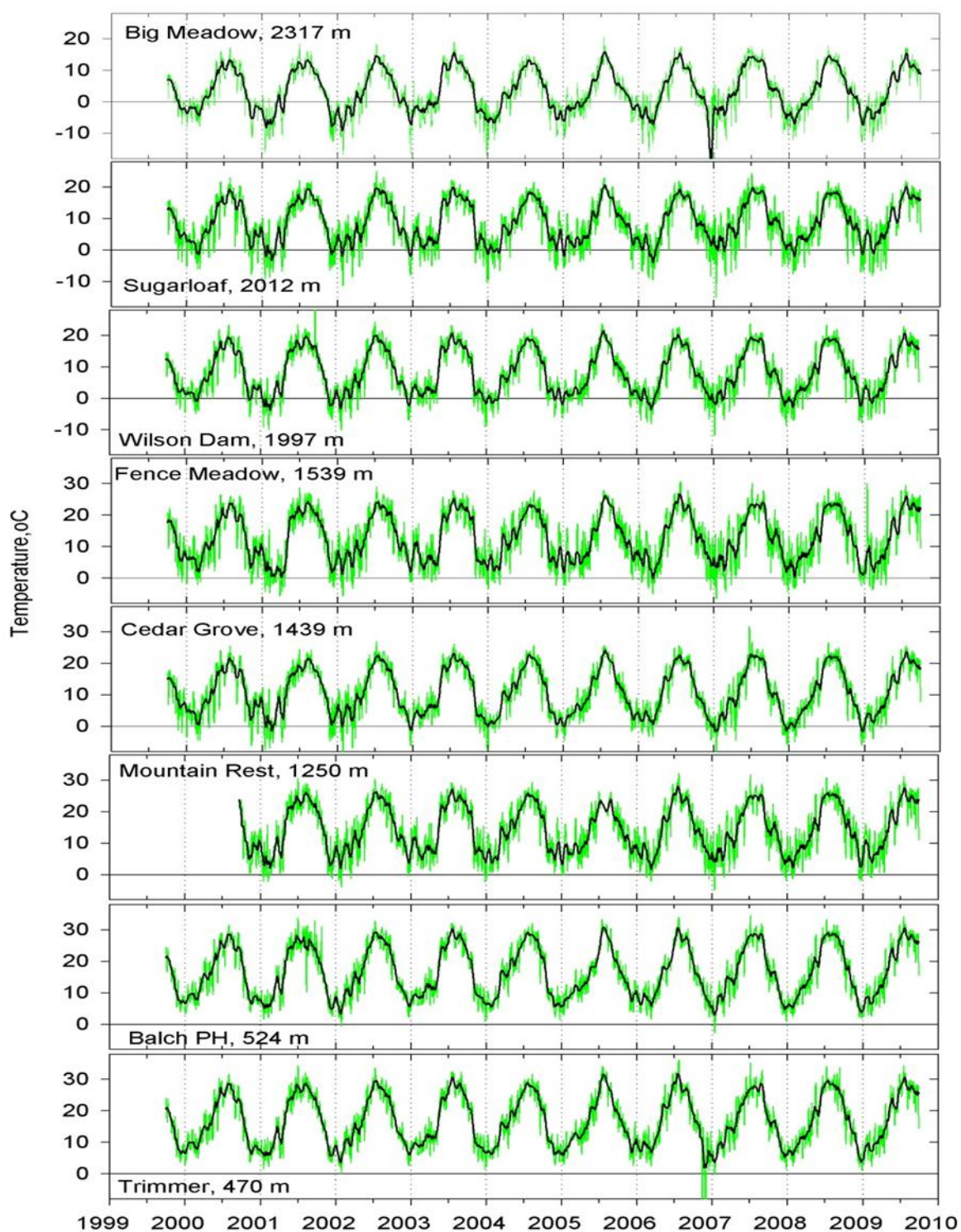


**Figure 5a (cont).** Daily average temperature for 16 stations in the San Joaquin river basin. Lines are daily and a 20-day running mean.



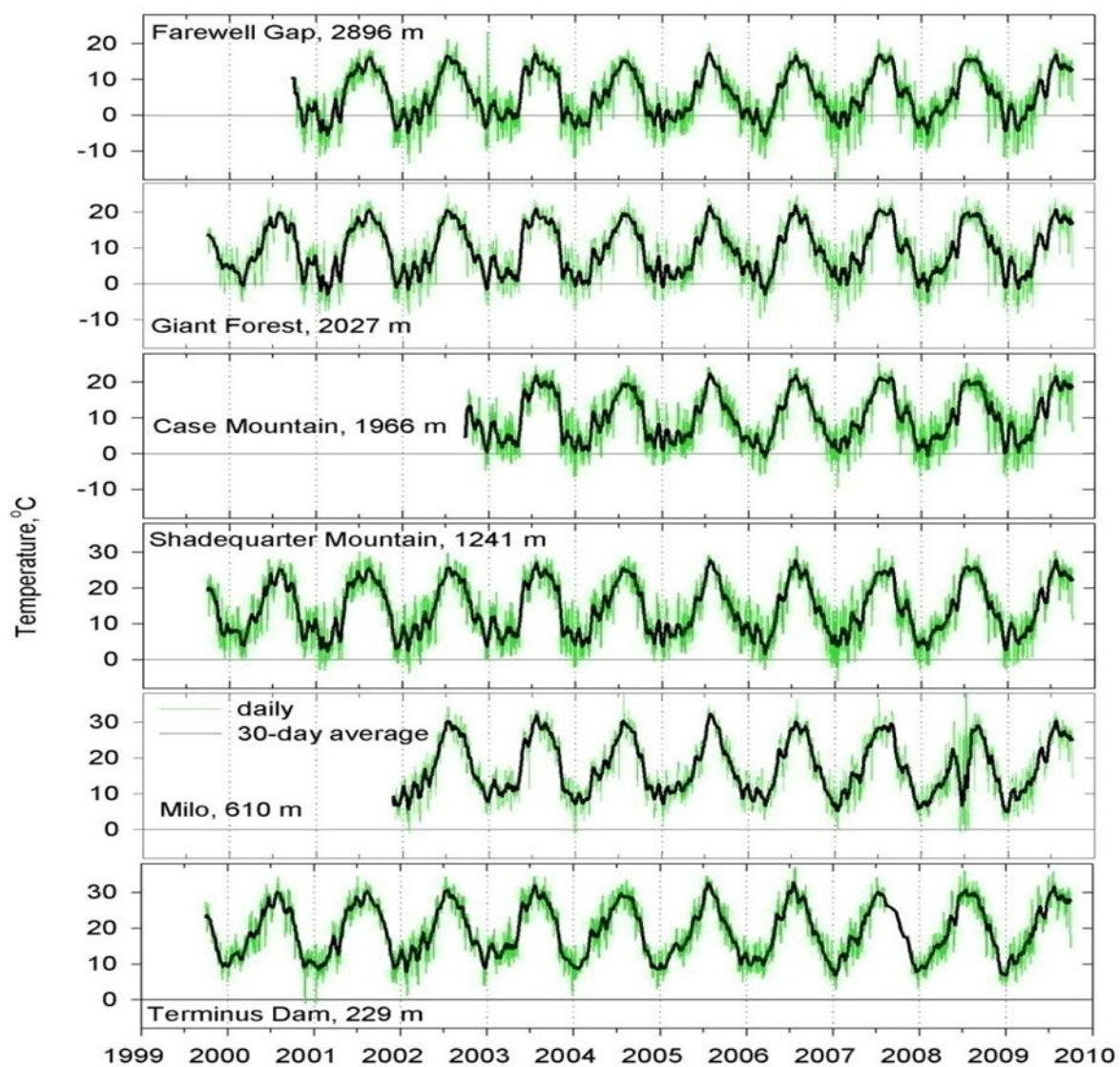


**Figure 5b.** Daily average temperature for 16 stations in the Kings river basin. Lines are daily and a 20-day running mean.

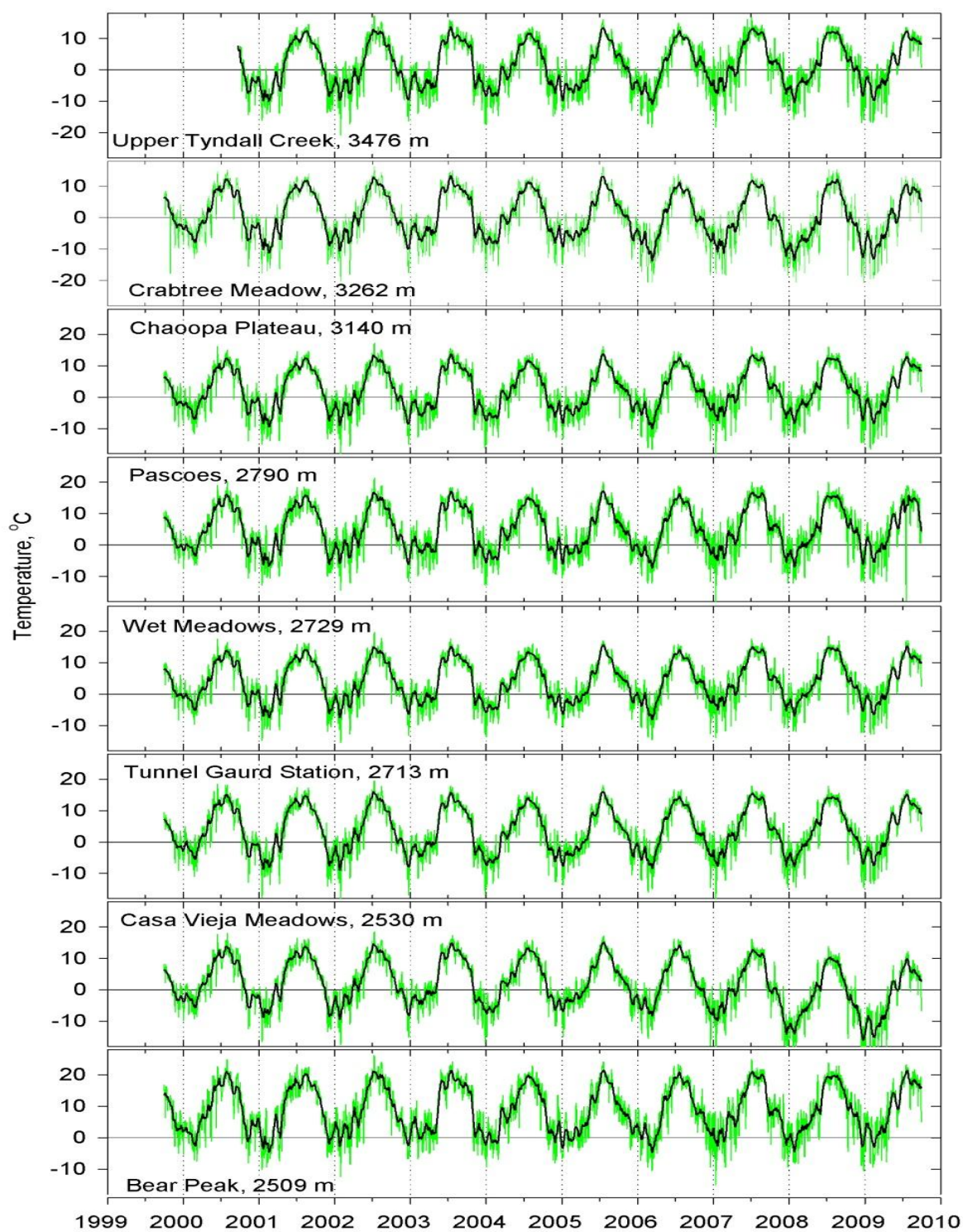


**Figure 5b (cont).** Daily average temperature for 16 stations in the Kings river basin. Lines are daily and a 20-day running mean.

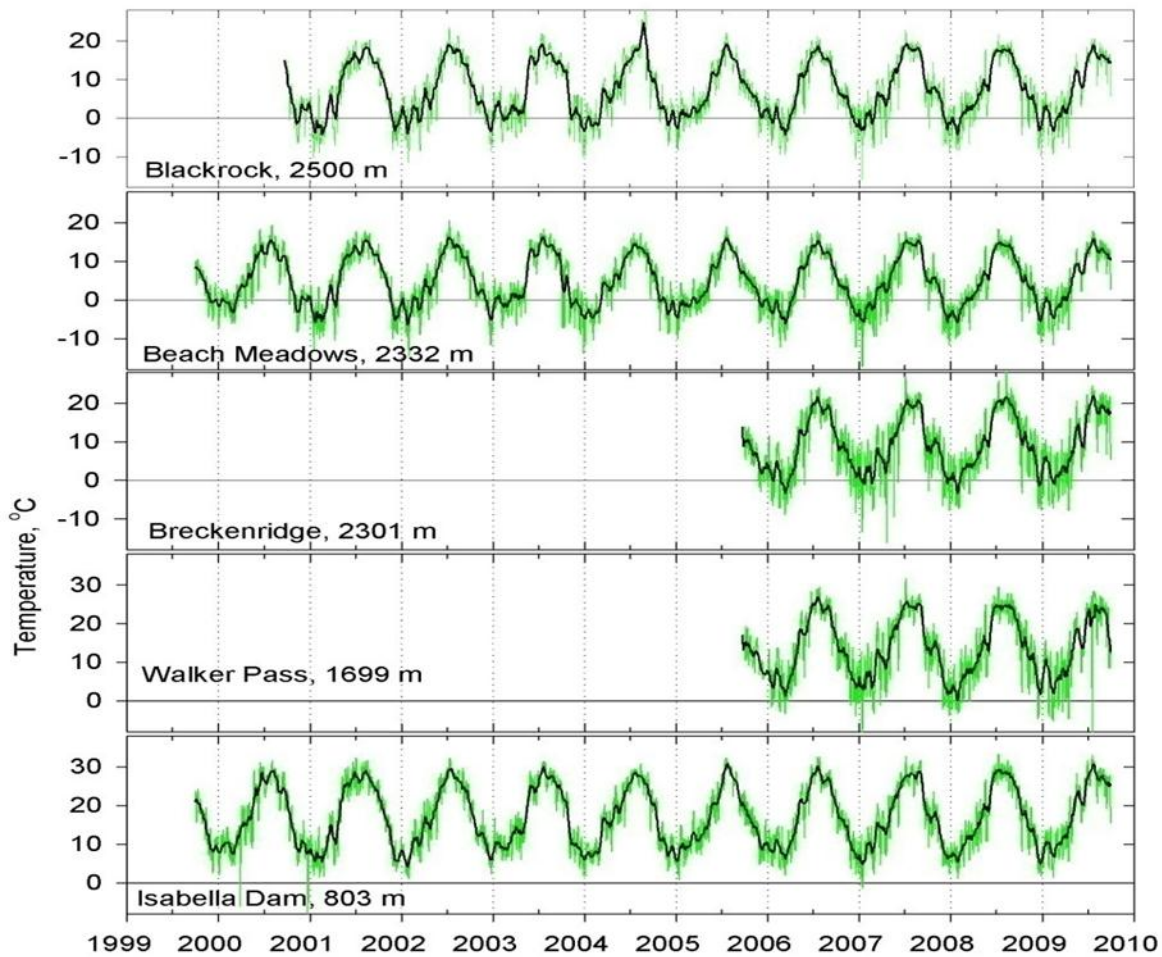




**Figure 5c.** Daily average temperature for 6 stations in the Kaweah river basin. Lines are daily and a 20-day running mean.



**Figure 5d.** Daily average temperature for 13 stations in the Kern river basin. Lines are daily and a 20-day running mean.



**Figure 5d (cont).** Daily average temperature for 13 stations in the Kern river basin. Lines are daily and a 20-day running mean.

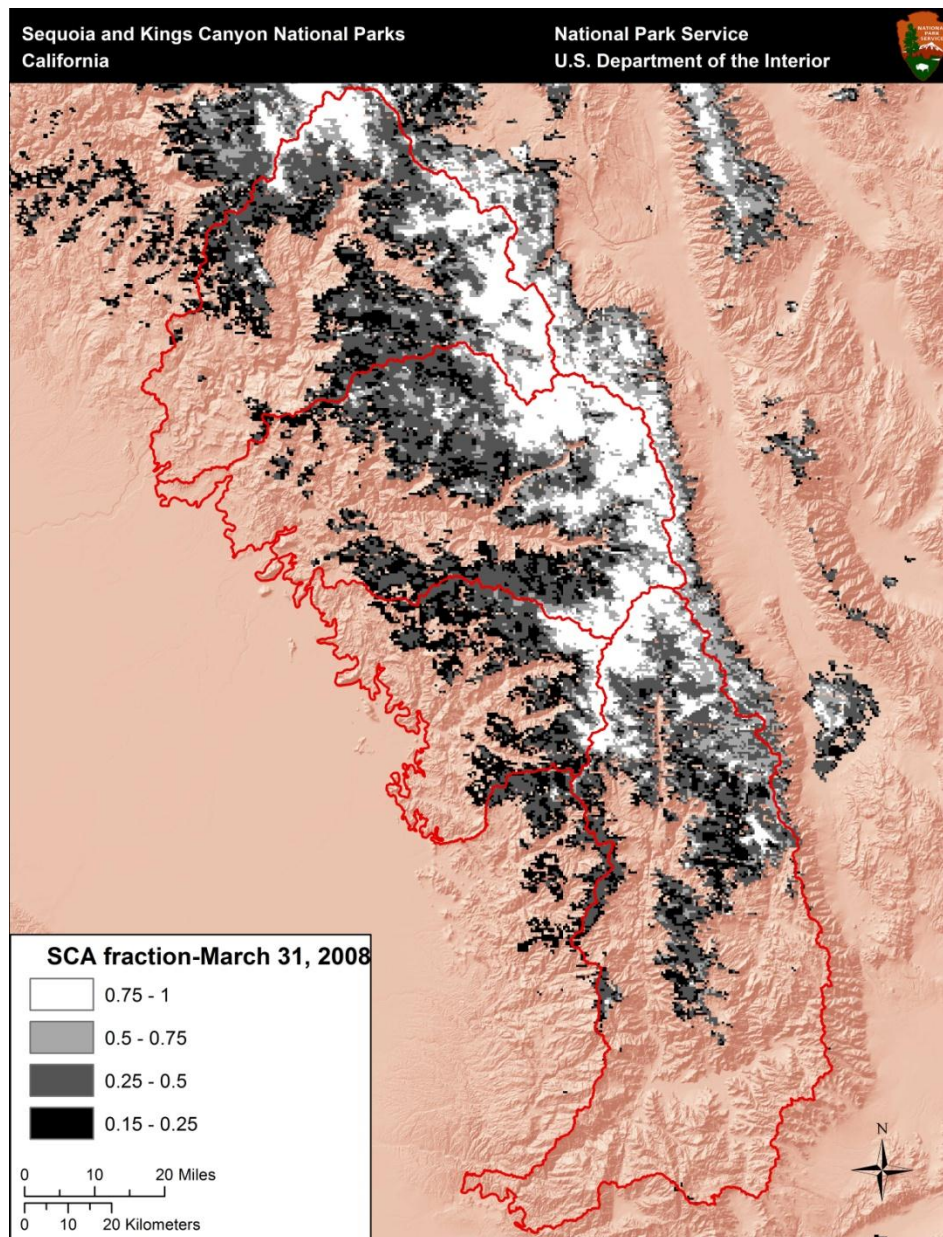
### Snow Covered Area

Snow covered area (SCA) from MODIS (Moderate Resolution Imaging Spectroradiometer) provides the only basin-wide coverages of snow, which is a dominant component of precipitation and runoff. Daily MODIS satellite retrievals give a relatively continuous record of SCA at a 500-m resolution (Figure 6). Cloud-free days were selected from the daily time series by visual inspection of both the SCA product and the MODIS cloud products. This resulted in SCA being available for approximately half of the days of interest in any given year. Daily SCA was then estimated using interpolation (3-parameter sigmoidal fit).

Daily SCA values for 2005, 2007, and 2008 were estimated from MODIS scenes using the MODSCAG algorithm (Painter et al., 2009). MODSCAG uses the MODIS surface-reflectance product (MOD09), sampled at a 500-m resolution in 7 spectral bands and corrected for atmospheric scattering (Kotchenova et al., 2006). The MODIS SCA product is retrieved from the satellite reflectances using a spectral-mixing model in which a set of end members (snow, rock, vegetation), present in different proportions in each pixel, is used to “unmix” a scene on a pixel-by-pixel basis. The algorithm uses the spectral information from MODIS to estimate subpixel



snow properties: fractional SCA, grain size and albedo (Painter et al., 2009). The daily MODIS snow-cover products were developed from the Terra satellite.



**Figure 6.** Fractional SCA for March 31, 2008. SCA values are binned into 4 fractions for ease of viewing (see legend). These values represent the fraction of the pixel that was detected by the satellite as being snow covered. Resolution of SCA pixels is 500 m. Red polygons are hydrologic basin boundaries.

### Snow Melt and Snow Water Equivalent

Snow water equivalent (SWE) is measured at 29 locations in the basins: 11 in the San Joaquin, 8 in the Kings, 2 in the Kaweah, and 8 in the Kern (Figure 1). SWE measured by the various water resources managers in the basins are primarily used for real-time, operational streamflow forecasts of seasonal runoff. These index sites (i.e. snow pillows) generally provide seasonal

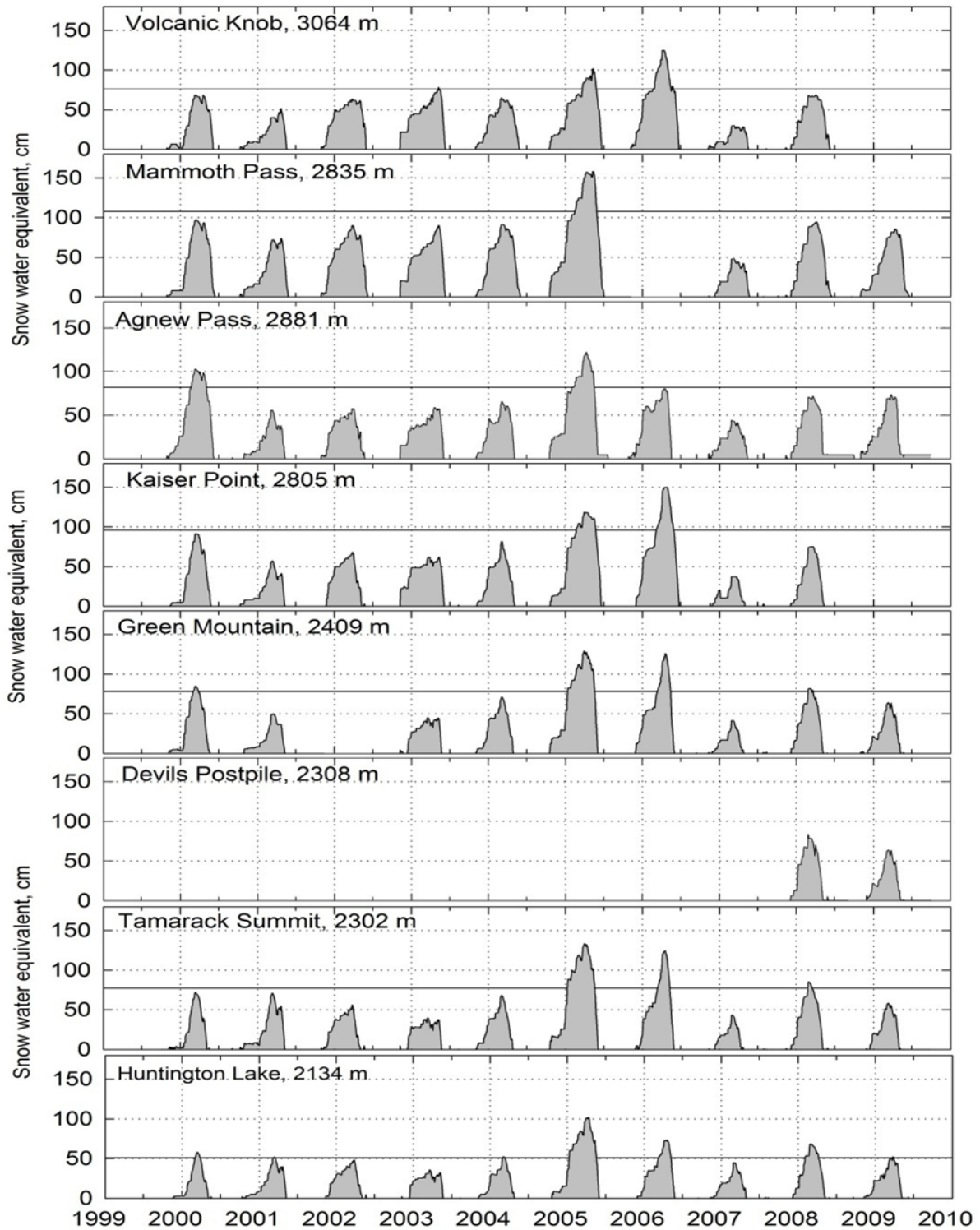
runoff forecasts of acceptable accuracy, they fail to provide spatially representative measures of SWE and do not capture the physiographic variability across a basin (Dressler et al., 2006; Molotch and Bales, 2005; Rice and Bales, 2010). Note that there is a lack of stations with consistent SWE data for the full 10-year period. The main records are shown on Figures 7a-7d. Note that despite a 1000-m elevation range for snow pillows on the San Joaquin and Kings basins, they do not show a consistent elevational difference in peak SWE reflecting in part the wide spatial variability in SWE within all elevation ranges. Only the lower-elevation stations in the Kaweah and Kern show significantly less SWE, likely due to a combination of less precipitation and a higher fraction of the precipitation falling as rain versus snow.

Findings from longer-term snow courses are also available; however those measurements are only monthly snapshots of snowpack water equivalent at 69 index sites (25 San Joaquin, 22 Kings, 5 Kaweah, and 17 Kern) in the four basins.

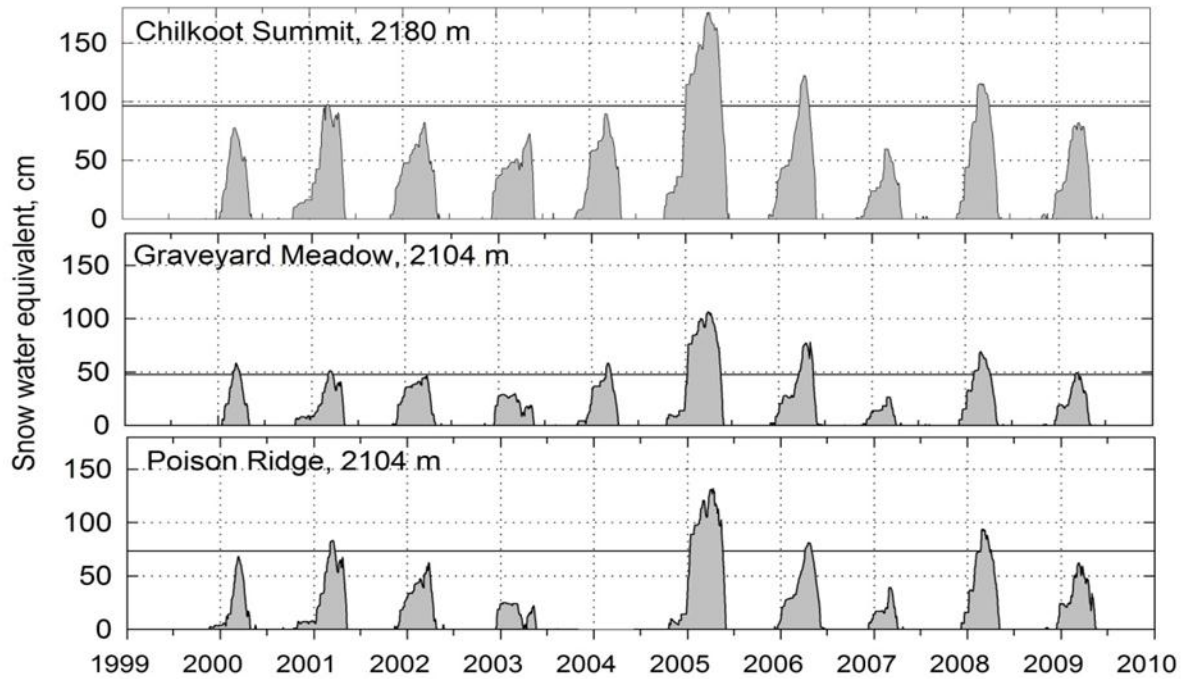
Daily snowmelt was estimated for each 300-m elevation band in each of the four basins analyzed based on the fractional SCA in that elevation range, and the daily temperature. The cumulative snowmelt was then summed back in time to get seasonal and annual total snowmelt (Cline et al., 1998; Liston, 1999; Molotch et al., 2004). This “depletion” approach assumes that the amount of snow that melted in a given area on a given day is equivalent to the net energy available for snowmelt on that area times the SCA for that area. Daily snowmelt was back-calculated using a temperature-index, degree-day equation rather than a full energy-balance model (Anderson, 1968; Granger and Male, 1978; Kuusisto, 1980):

$$M = a(T_a - T_b) \quad (1)$$

where  $M$  is the daily snowmelt,  $a$  is a degree-day coefficient ( $\text{m deg}^{-1} \text{ day}^{-1}$ ),  $T_a$  is average daily temperature,  $T_b = 0^\circ\text{C}$ ; when  $T_a < T_b$ , no melt occurs (Kustas et al., 1994). This depletion-method calculation was driven by daily average temperature, which was calculated across the elevation range of the study area using a ground-surface lapse rate that was estimated from stations in the study area (see below). The approach provides an index of the average energy flux, but does not explicitly consider the individual fluxes and controlling factors that influence snowmelt (e.g. solar radiation, albedo, topography, turbulent energy exchanges). Daily data from snow-pillow sites were used to estimate coefficients for the degree-day calculation. Since temperature and SWE vary from year to year, producing different inter-annual daily melt rates, a daily degree-day coefficient for each year was calculated for the San Joaquin, Kings, Kaweah, and Kern River basins using the co-located snow-pillow and temperature measurements. Daily average temperature and SWE were calculated from the hourly CDEC data.



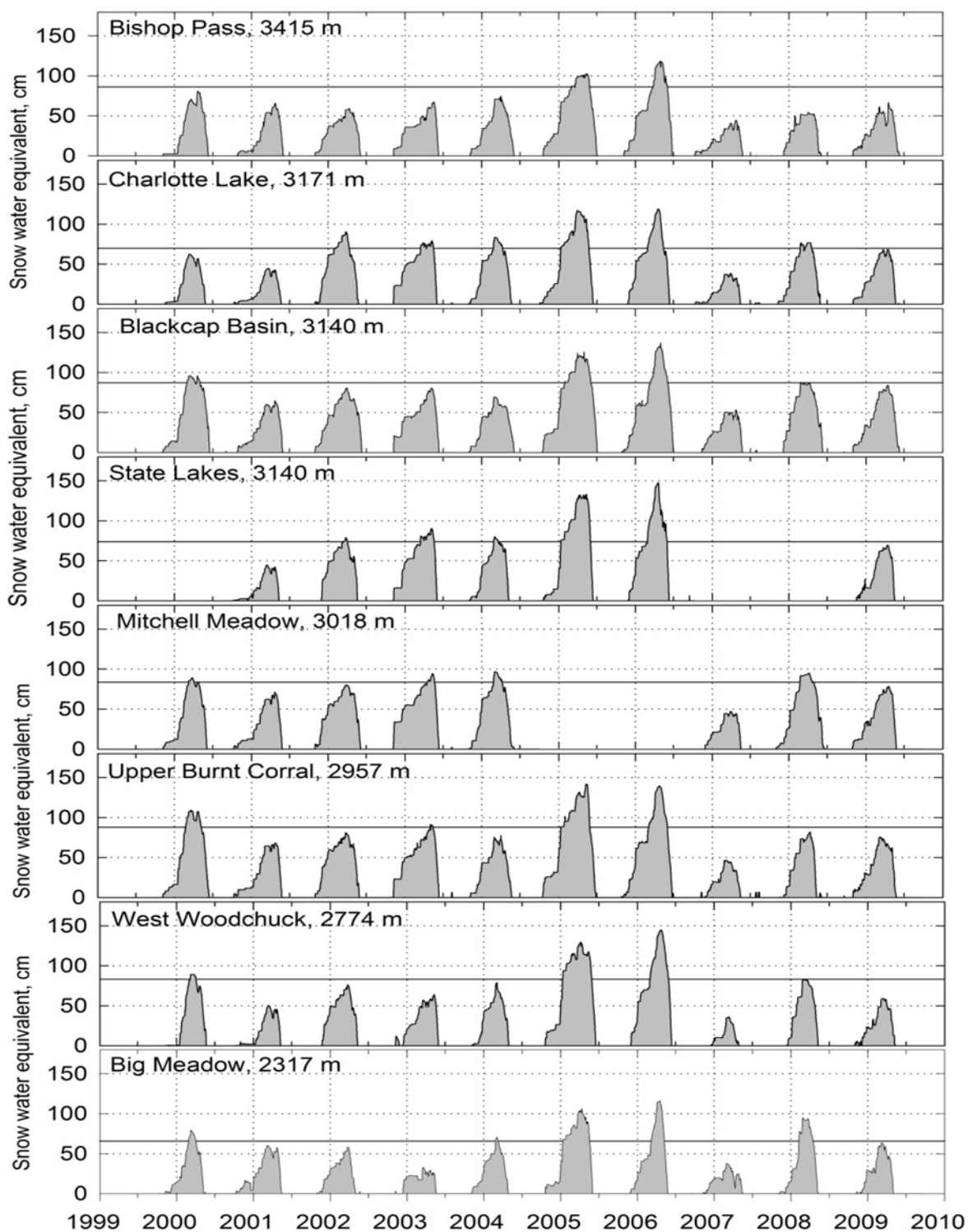
**Figure 7a.** SWE for 11 pillows in the San Joaquin basin. Note that some years have missing data (not zero values).



**Figure 7a (cont).** SWE for 11 pillows in the San Joaquin basin. Note that some years have missing data (not zero values).

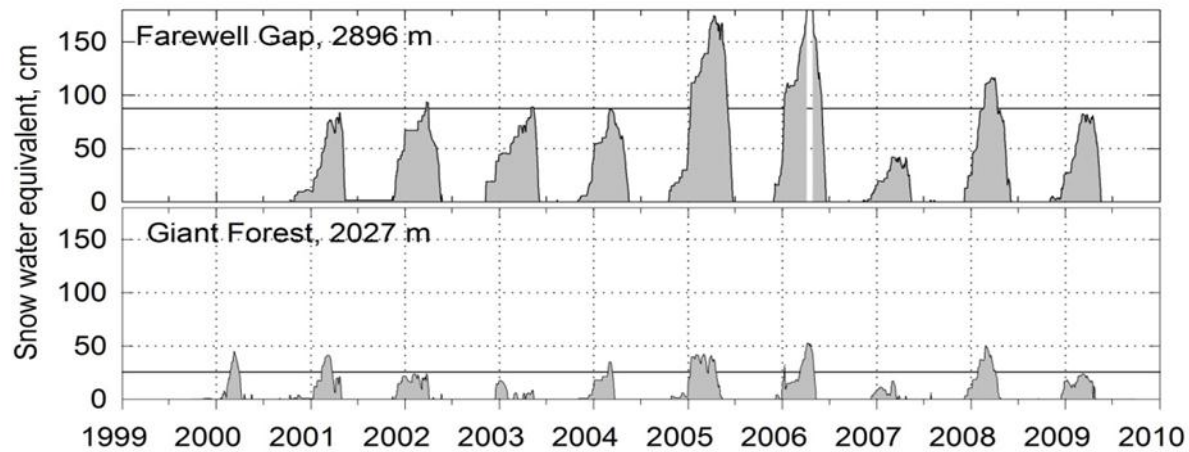
Values of the degree-day coefficient ( $a$ ) were estimated from data at each of the snow-pillow sites in the basins by dividing the amount of daily melt indicated by the snow pillow by the daily degree-day quantity. Days with near-zero melt and days with very high melt but near-zero degree days were not used. There was no consistent or statistically significant seasonal trend to the degree-day coefficient for any of the sites. Some sites had a large degree of inter-annual and intra-annual variability, with others showing much less scatter, but did increase over the ablation season (Figure 8). Each of the 4 basins had different degree-day coefficients, reflecting the differences across latitude, e.g., north versus south, and in part because, the number of snow pillow and co-located temperature stations were sufficient to generate basin specific degree-day coefficients (Figure 9). Since the snow-covered part of the basin extends over more than 2000 m of elevation, daily degree-day coefficients were calculated for snow pillows and co-located temperature stations. Snow pillows and temperature stations were located in for 4 of the 9 elevations bands between 2100-3300 m for the San Joaquin River basin, 5 of the 9 elevations bands between 2100-3600 m in the Kings and Kern River basins, and 3 of the 8 elevations bands between 2100-3000 m in the Kaweah River basin. Stations were grouped by elevation band and a linear trend fit to the values. In addition, in order to reduce the effects of possible site-specific differences (i.e. local shading, vegetation) snow pillow stations within  $\pm 80$  m of an elevation band were included to increase the sample size and reduce the bias. It should be noted that no station data were available above the 3000-3300 m elevation band in the San Joaquin, 3300-3600 m elevation band in the Kings and Kern, and 2700-3000 m elevation band in the Kaweah River basins. Above and below elevation bands where station data are available, daily coefficients from the adjacent elevation band were applied. It should be noted that not all of sites had usable data for estimating the coefficients for each year, and that the slopes to the fitted lines for the individual years were not statistically significant (Figure 9).





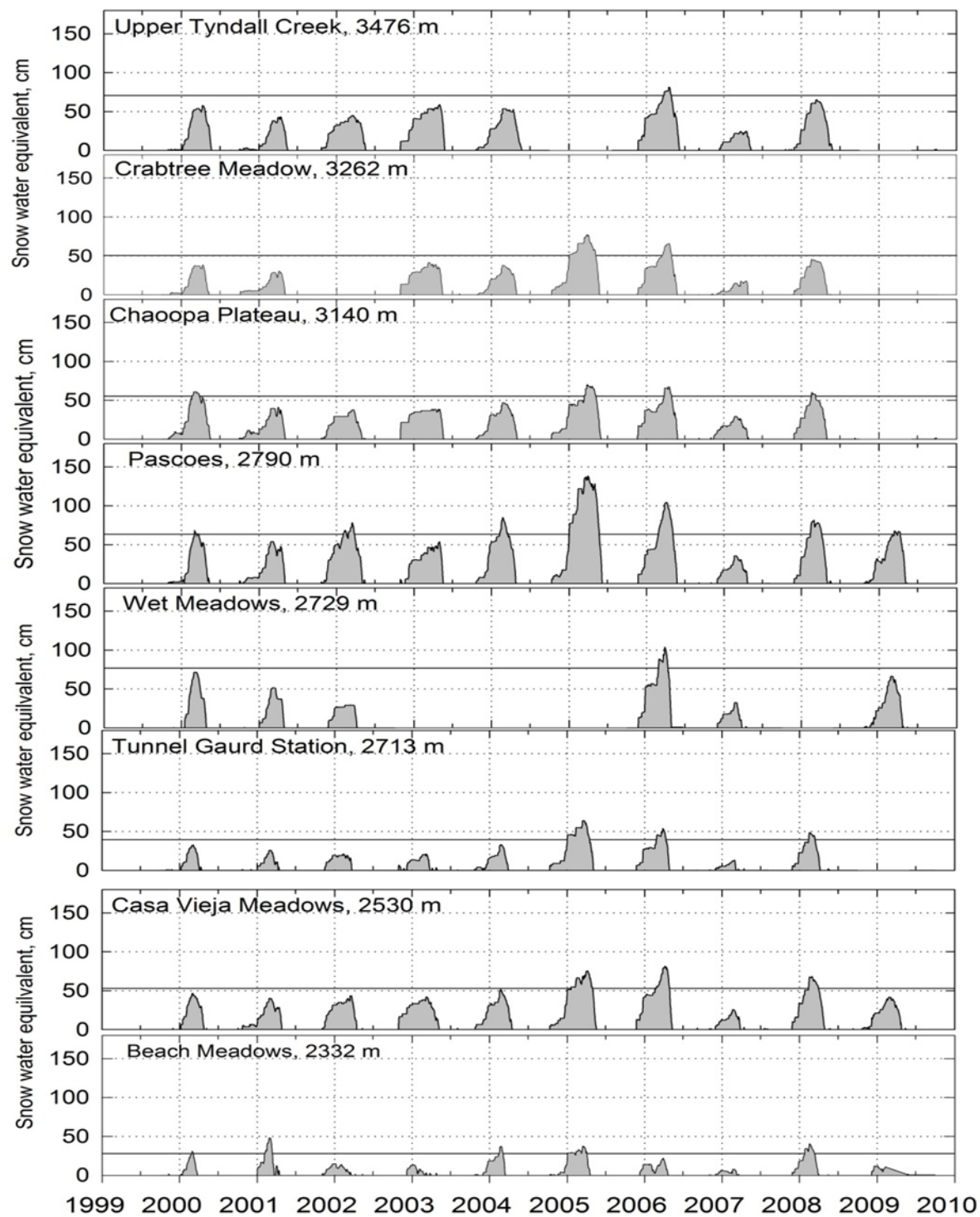
**Figure 7b.** SWE for 8 pillows in the Kings basin. Note that some years have missing data (not zero values).



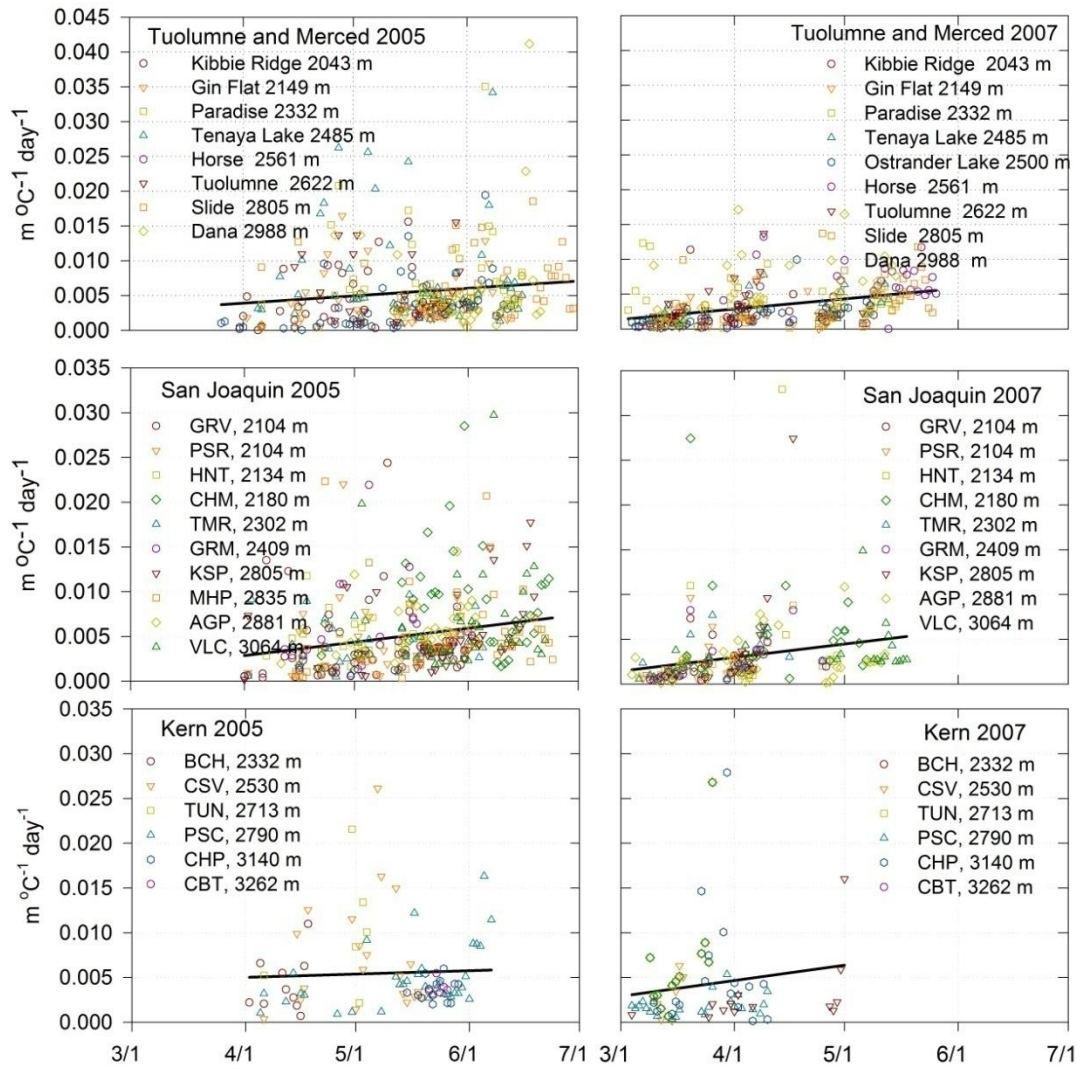


**Figure 7c.** SWE for 2 pillows in the Kaweah basin. Note that some years have missing data (not zero values).

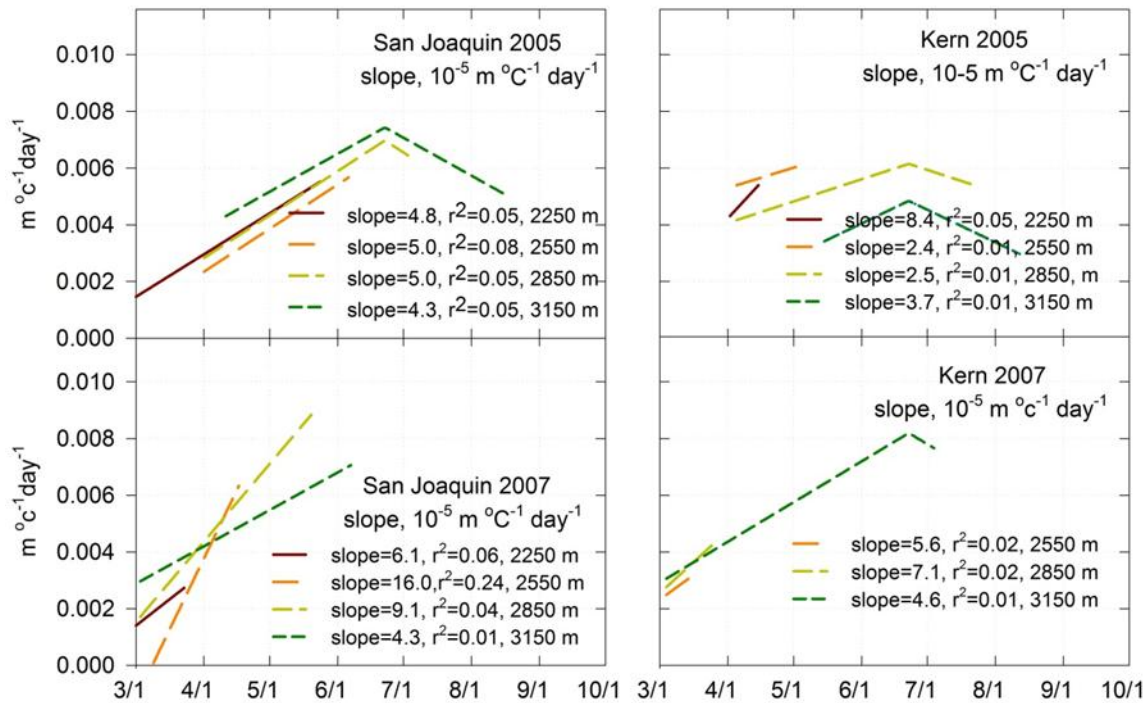
This method of parameterizing the daily degree-day coefficient was similar to the methods developed by Martinec (1960) and Anderson (1968) and used in such semi-distributed runoff models as the Snowmelt Runoff Model (SRM) and Snow-17. The SRM uses a 3-5 day moving average of the daily degree-day coefficients to provide a more consistent coefficient, in addition to developing coefficients based on elevation bands of 500 m when areal computations of snowmelt are required. Anderson (1968) developed a sinusoidal relationship from 5-years of parameterizing the daily degree-day coefficients at the Central Sierra Snow Lab near Donner Pass, California and by knowing the maximum and minimum degree-day coefficient on June 21 and December 21, respectively, the daily degree day coefficient throughout the year can be computed from the sinusoidal relationship.



**Figure 7d.** SWE for 2 pillows in the Kern basin. Note that some years have missing data (not zero values).



**Figure 8.** Degree-day coefficients for snow-pillow sites in the San Joaquin and Kern River basins for a wet (2005) and a dry (2007) years.



**Figure 9.** Elevation-dependent degree-day coefficients across 4 of 9 elevation bands in a wet (2005) and a dry (2007) year. It should be noted that only 3 elevation band are shown in the Kern River basin for 2007 due to insignificant snowmelt at elevations below the 2550 m elevation band. Values are shown for period of snowmelt at each elevation.

## Precipitation

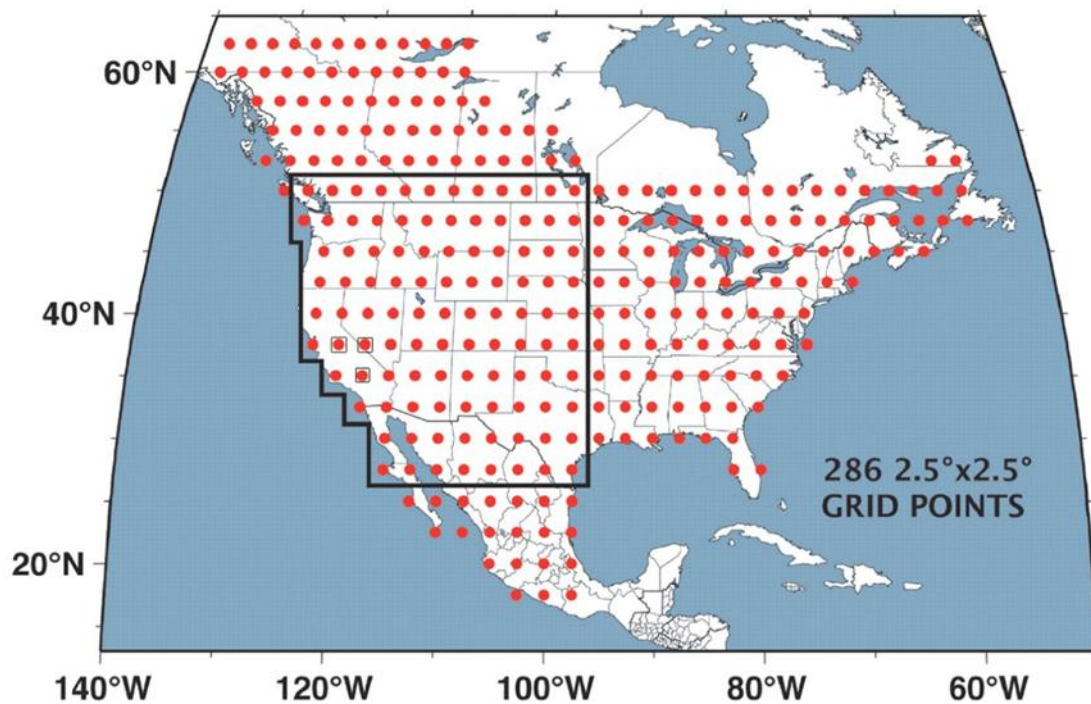
Precipitation is measured at 59 locations in the basins, however, due to inconsistent data quality and data dropouts throughout the period of record, 31 precipitation stations were used: 15 in the San Joaquin, 6 in the Kings, 5 in the Kaweah, and 5 in the Kern (Figure 1). For example RAWS stations use non-heated tipping buckets, and during the winter due to significant snowfall, fail to record reliable measurements. Note that there is a lack of stations with consistent precipitation data for the full 10-year period that is the main focus of this analysis.

PRISM precipitation data for the 800 m normals (1971-2000) and the 4 km monthly (2000-2009) were downloaded from (<http://www.prism.oregonstate.edu/>).

## Paleoclimate

In both the paleoclimate and instrumental record, droughts are among the events of greatest concern to resource managers and other stakeholders whose life and livelihood depend on the availability of fresh water and the health of mountain ecosystems. The definitive reconstruction of drought for the United States by Cook, et al. (2004) is a gridded 2.5o PDSI data set covering most of North America (286 points) (Figure 10). PDSI has been the most commonly used drought index in the United States and was developed to measure intensity, duration, and spatial extent of drought. Although precipitation is a major component of the PDSI (Palmer Drought Severity Index), the values are derived using additional parameters such as air temperature and local soil moisture, along with prior values of these measures. Values range from -6.0 (extreme drought) to +6.0 (extreme wet conditions). The Cook et al. (2004) PDSI reconstructions are

based on a network of 835 tree-ring chronologies (602 western), and supersede those based on single sets of chronologies like that of Graumlich (1993) for the Sierra Nevada. The temporal coverage extends back nearly 2000 years for some locations. For the Sierra Nevada and western Great Basin, the three grid points highlighted on Figure 10 are of potential relevance to SEKI. It should be noted, however, that given the spatial coarseness of the product, these are broad regional values that cover a very topographically and climatically diverse region. There is likely considerable spatial variability in interannual climate response within these grid points.



**Figure 10.** North American grid developed by Cook et al., 2004. The 3 grid points considered in the current analysis are enclosed in small squares.





## Reference conditions

### Temperature

The density, quality and record length of daily average temperatures for the Kaweah basin (Figure 5c) are typical of those available for the Parks and the 4 river basins. Several records go back about ten years with few gaps. Note the relatively consistent annual cycles among the 6 stations, with about a 20°C summer-winter difference in average temperature. Because most stations do not have long records, model products based on broad interpolation with lower-elevation stations are often used for earlier periods. The appendix on Landscape Context (Appendix 1) has provided and documented change from a 30-year mean of 1910-1940, and how temperature and precipitation has changed using a contemporary mean from 1970-2000 based on the available PRISM temperature and precipitation product. Caution is urged, however, in using this sort of broad, coarse and temporally averaged dataset to infer local changes. There is also significant within-grid and within region variability owing to small-scale topography. For example, daily minimum and maximum differences in SEKI range from 11.6 to 14.5°C, possibly reflecting different amounts of nighttime cold-air drainage at Lodgepole and Ash Mountain versus Grant Grove and Giant Forest (Figure 4)

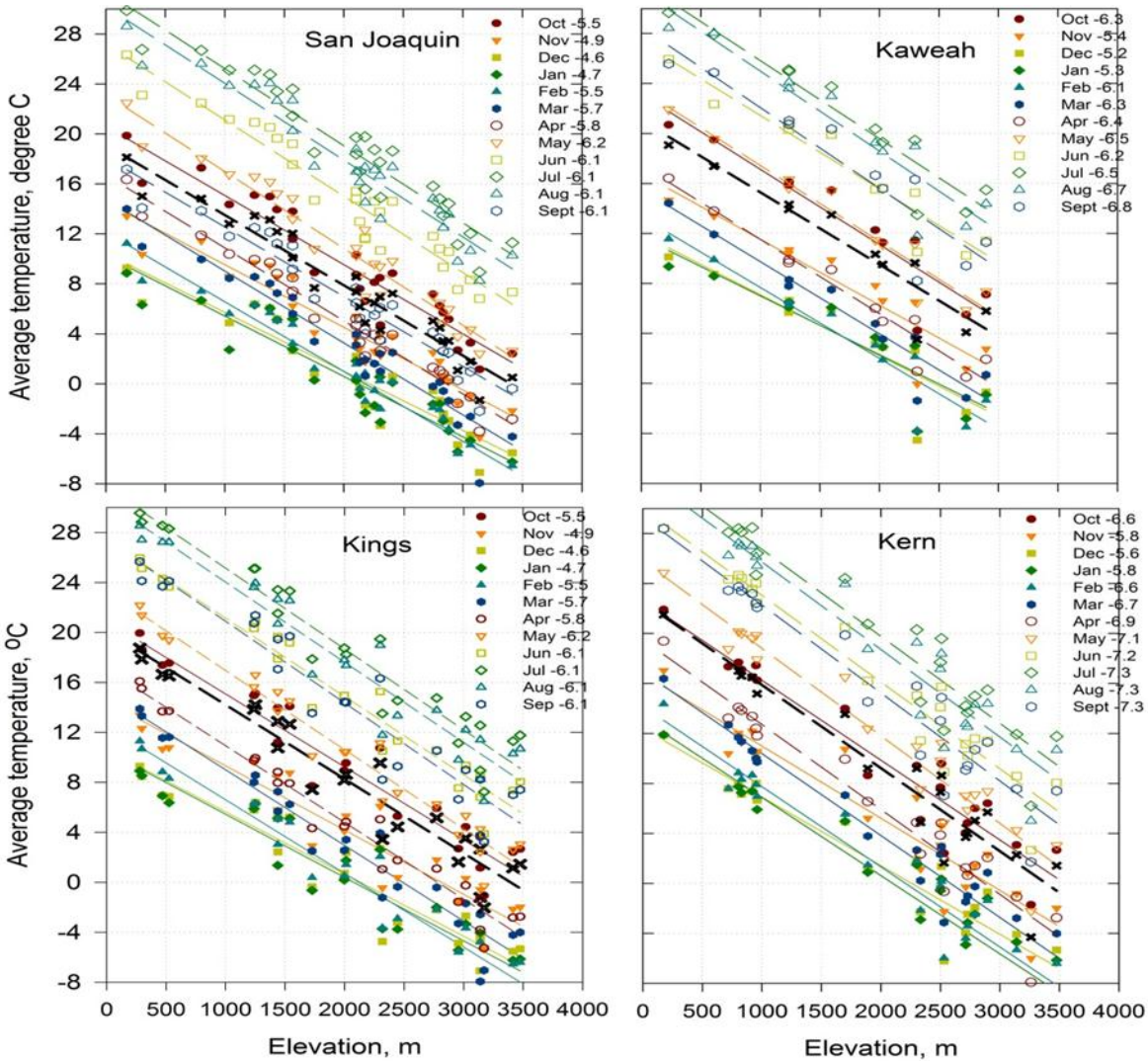
Twenty-two stations in the San Joaquin, 20 stations in the Kings, eight stations in the Kaweah, and 20 stations in the Kern were used to estimate daily, annual average, and monthly ground-level “lapse rates” for the ten-year period 2000-2009 (Figure 11). Monthly values ranged -4.6 to -7.3°C per 1000 m elevation; with an annual average of -5.6, -5.9, 5.8, and -6.7°C per 1000 m for San Joaquin, Kings, Kaweah and Kern River basins, respectively. Values are lowest in winter and highest in summer. For the main snow-accumulation season, December through March, the average ranges from -5.1 to -6.2°C per 1000 m; for the main snowmelt season, February through June, the average ranges from -5.9 to -6.9 °C per 1000 m, corresponding to approximately a 1.7 to 2°C change in average temperature for each 300-m (1000 ft) elevation change.

For comparison, the annual average temperature change with elevation in the gridded PRISM data over the same elevation range is somewhat lower, about -5.3 to -6.1°C per 1000 m (1971–2000 mean annual) and -5.0 to -5.9 per 1000 m (2000-2009), as shown on Figure 12. That is, the slope of the temperature-elevation relation is less steep for the PRISM data than for the stations used in the current analysis. Note that PRISM data are on an 800-m grid for the 1971-2000 normals and 4-km grid for the 2000-2009 monthly temperature data, and that elevation varies by at least 300 m over this distance in many snow-producing grid cells. We thus consider the station data to provide a better estimate of ground-level lapse rate than the gridded data. In addition, it should be noted that PRISM products may not include all available mountain temperature records, and should be checked versus local temperature records.

### Snow Covered Area

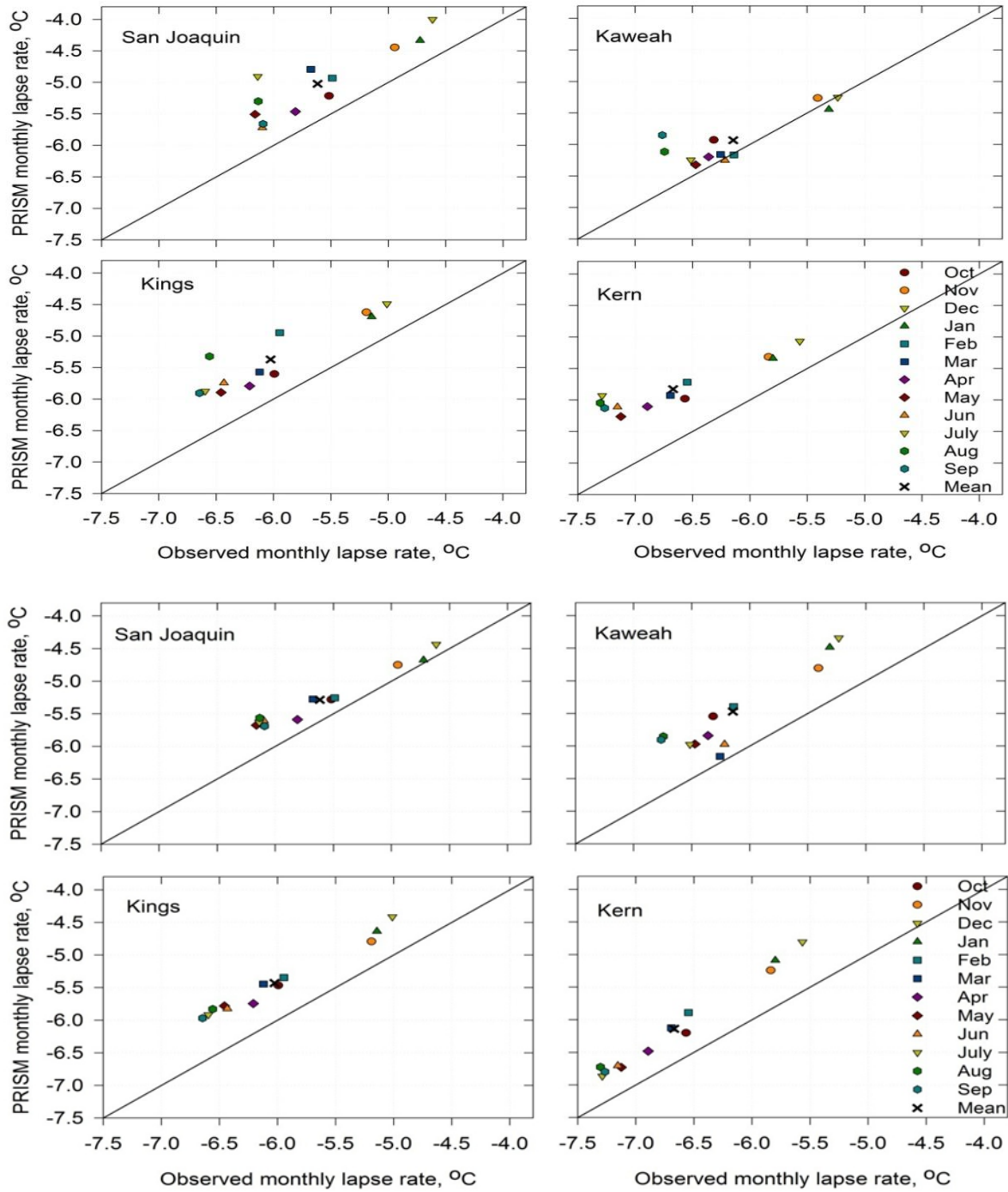
We restricted this analysis to seasonally snow-covered areas, which start at the rain/snow transition of 1500 m, representing 72% (3199 km<sup>2</sup>) of the San Joaquin, 66% (3172 km<sup>2</sup>) of the Kings, 34% (835 km<sup>2</sup>) of the Kaweah, and 75% (4628 km<sup>2</sup>) of the Kern River basins. Using topographic data from the Shuttle Radar Topography Mission at 30-m spatial sampling (Farr et al., 2007), the study area was partitioned into eight (Kaweah) or nine (San Joaquin, Kings, Kern) elevation bands of 300-m increments, beginning at 1500 m and extending to greater than 3900 m.



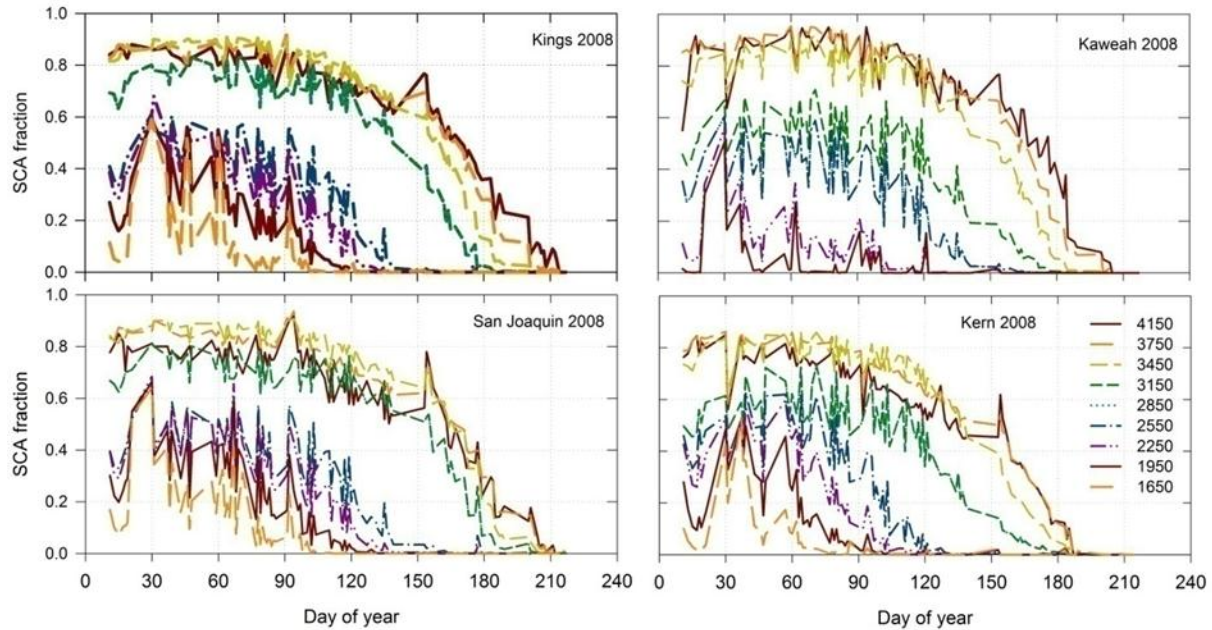


**Figure 11.** Monthly and annual average temperatures and lapse rates in the project area. Annual average marked by x and fit by heavy dashed line (slope -6.8°C per 1000 m). Slopes of the individual monthly fitted lines are given in the legend.

SCA results are presented in 300-m elevation bands for 2005, 2007 and 2008. SCA shows remarkable consistency across the four watersheds, with winter values near 1.0 at the highest elevations, and lower SCA values in successively lower elevation bands. This is illustrated for 2008 on Figure 13. In this figure, the average SCA values were averaged across all pixels in the 300-m elevation bands centered on the elevation values given in the legend. Since snow was only present in the 1350-m band in a few scenes, and values were generally near zero, i.e., its contribution as a snow producing region to the total amount was small. The analysis that follows used the nine (eight in the Kaweah) bands from 1650 to 4150 m. The area above 3900 m in the San Joaquin, Kings, and Kern was still relatively small (less than 0.5%) when compared to the total basin area.



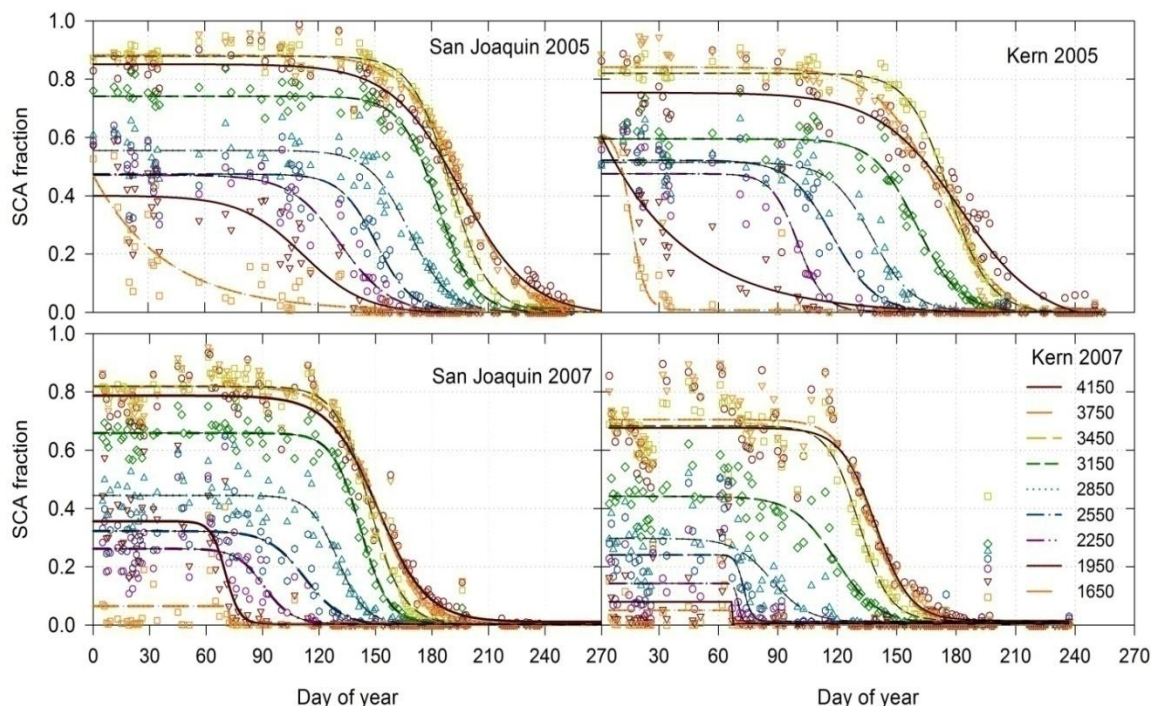
**Figure 12.** Comparison of the mean monthly lapse rates derived from station data (2000-2009) versus PRISM for: a) the same period at a 4-km grid resolution, and b). 1971-2000 normals) at an 800-m grid resolution. The solid line is a 1:1 line.



**Figure 13.** SCA time series for 2008 for the four main watersheds draining from Sequoia-Kings Canyon National Parks by elevation band (legend in m).

The San Joaquin and Kern river basins illustrate SCA patterns at the higher elevations in a wet year (2005) and dry year (2007) (Figure 14). In the San Joaquin River basin for 2005, representative of the northern portion of the Parks, SCA values reach 0.85-0.90 in winter at the highest elevations (3450, 3750, and 4150 m elevation bands), drop slightly in the next 300 m (3150 m elevation band), and then drop off more steeply over the next 300 m (Figure 14). The drier year 2007 shows a similar trend, but with less snow, with SCA values reaching 0.80 at the highest elevations, dropping slightly to 0.75 over the next 300 m (3150 m), and again dropping off steeply at the lower elevations (<3000 m). In Kern River basin, representative of the southern part of the Parks, SCA values in 2005 reached 0.75-0.85 in winter at the highest three elevations (3450, 3750, and 4150 m), drop to 0.48-0.6 in the next 4 elevations (2250, 2550, 2850, and 3150m), and drop off more steeply over the next 900 m (Figure 14). In 2007 the highest 3 elevations have a lower SCA value (0.70) across all 3 bands, with significantly lower SCA at the mid- to lower- elevations. Both 2005 and 2007 exhibit one main spring melt, with snow cover declining from relatively constant winter values to zero over a period of 2-3 months, except at the lowest elevations where melt was over a period of 1-2 months. In the analysis that follows, 2005 and 2007 are used as indicative of the snow-producing elevations during a wet and dry year, respectively. They are used as well as representative differences between the northern and southern portion of the NRCA study region, with the San Joaquin and the Kern used to illustrate SCA in the Northern and Southern regions.



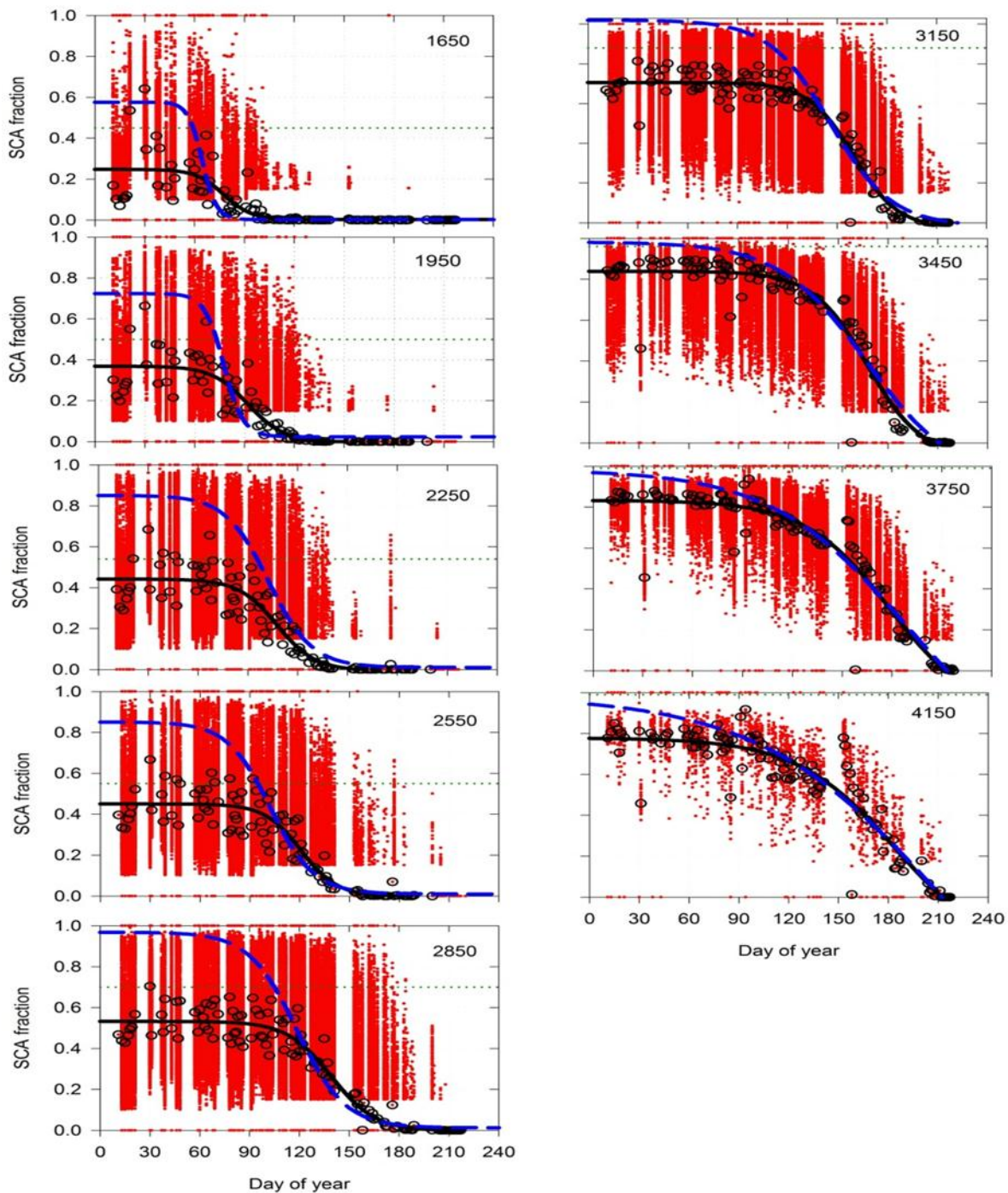


**Figure 14.** San Joaquin and Kern River basin SCA, with a 4-parameter sigmoidal fit by elevation band (legend in m).

One main factor limiting the accuracy of the satellite-derived SCA values is canopy cover, or the fraction of the land surface that is beneath tree canopies that limit satellite observation of the ground. That is, the satellite can only detect snow that is in between tree canopies, not under tree canopies. Thus corrections were made for canopy cover. Full-coverage, recent data for canopy closures for these bands were from the National Landcover Data (<http://www.mrlc.gov/>) and Sequoia and Kings Canyon National Park Vegetation map (<http://ninfo.nps.gov/Geospatial.mvc/Welcome>). Corrections for SCA under canopy were made empirically by examining the range of SCA and canopy cover across elevation bands for individual pixels in the elevation band. A second issue with the SCA estimates is the threshold of 0.15 for SCA detection by the MODSCAG algorithm, limiting detection of residual SCA as pixels melt out. It should be noted that complex topography also limits full viewing of SCA for off-nadir viewing angles.

Adjustments of 0.15-0.30 were made to the SCA fraction in each elevation band, and the resulting data fit to a 4-parameter sigmoidal curve (Figure 15). On this figure, the SCA values for each pixel and each day in 2008 are shown as a point; note that 2008 was about average SCA for the period of record (1999-present). The average line for the pixels is shown as a sigmoidal fit to the average SCA, as well as the sigmoidal fit to the SCA after adjustment. The corrected sigmoidal fits were used for subsequent analyses. Though ground data for a formal evaluation of the correction are not available, estimates made when the higher elevations are essentially completely snow covered suggest that uncertainty is on the order of 5%. That is, complete snowcover represents coverage of 95-100%, depending on the presence of slopes too steep to hold snow and trees. The same magnitude of vegetation and threshold corrections were applied to each year, up to a SCA fraction of 1.0. Corrections were larger in the mid elevations, which

have greater canopy cover. Note that while canopy effects will be larger at low satellite viewing angles, that angle effect was not accounted for in the analysis.

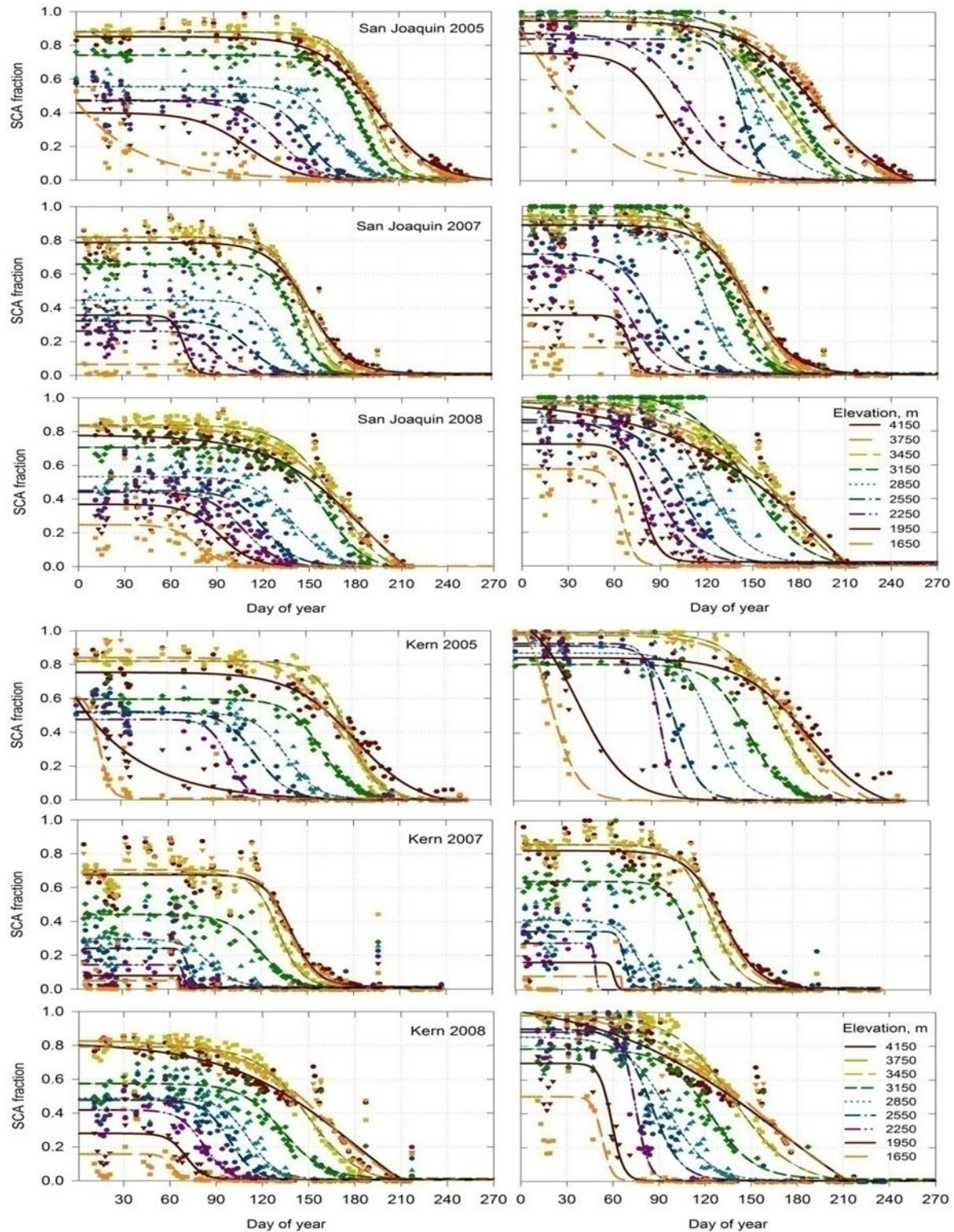


**Figure 15.** San Joaquin River basin SCA, 2008, presented in 300-m elevation bands (meters), centered on the elevations indicated in each panel. Each closed dot is the SCA for an individual pixel, with the open circles being the daily average and the solid line the to a 4-parameter sigmoidal curve fit. The dashed line is the adjusted fit.

Three years further illustrate the magnitude of the corrected SCA: i) the relatively wet year 2005, ii) the relatively dry year 2007 and iii) the more-average year 2008 (Figure 16). Note that SCA in the San Joaquin River basin at highest elevation bands was corrected to near 100%, with values in lower elevation bands receiving adjustments of similar magnitude.

Note that in the lowest elevation bands, snowmelt occurs relatively quickly in winter. SCA in the more-southern Kern River basin is lower than that in the more-northern San Joaquin River basin, and snowmelt was earlier. Only in the wet year (2005) was there significant snow-covered area in the lowest elevation band, 1,650 m. In the dry year (2007) there was also little snow-covered area in the 1950 and 2250 elevation bands. Note that the 4 river basins have mean elevations in the 2200-2600 m range, and with 10-30% or the area above 3000 m (Figure 17).





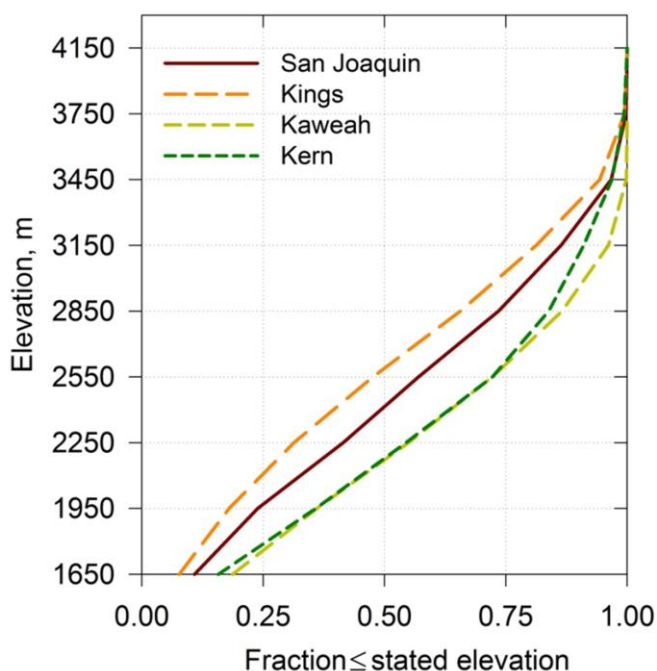
**Figure 16.** Comparison of the raw (left panels) and adjusted (right panels) SCA for a wet (2005), dry (2007), and average (2008) year for the: a) San Joaquin River basin and b) Kern River basin, by elevation band (legend in m).



## Snow Melt and Snow Water Equivalent

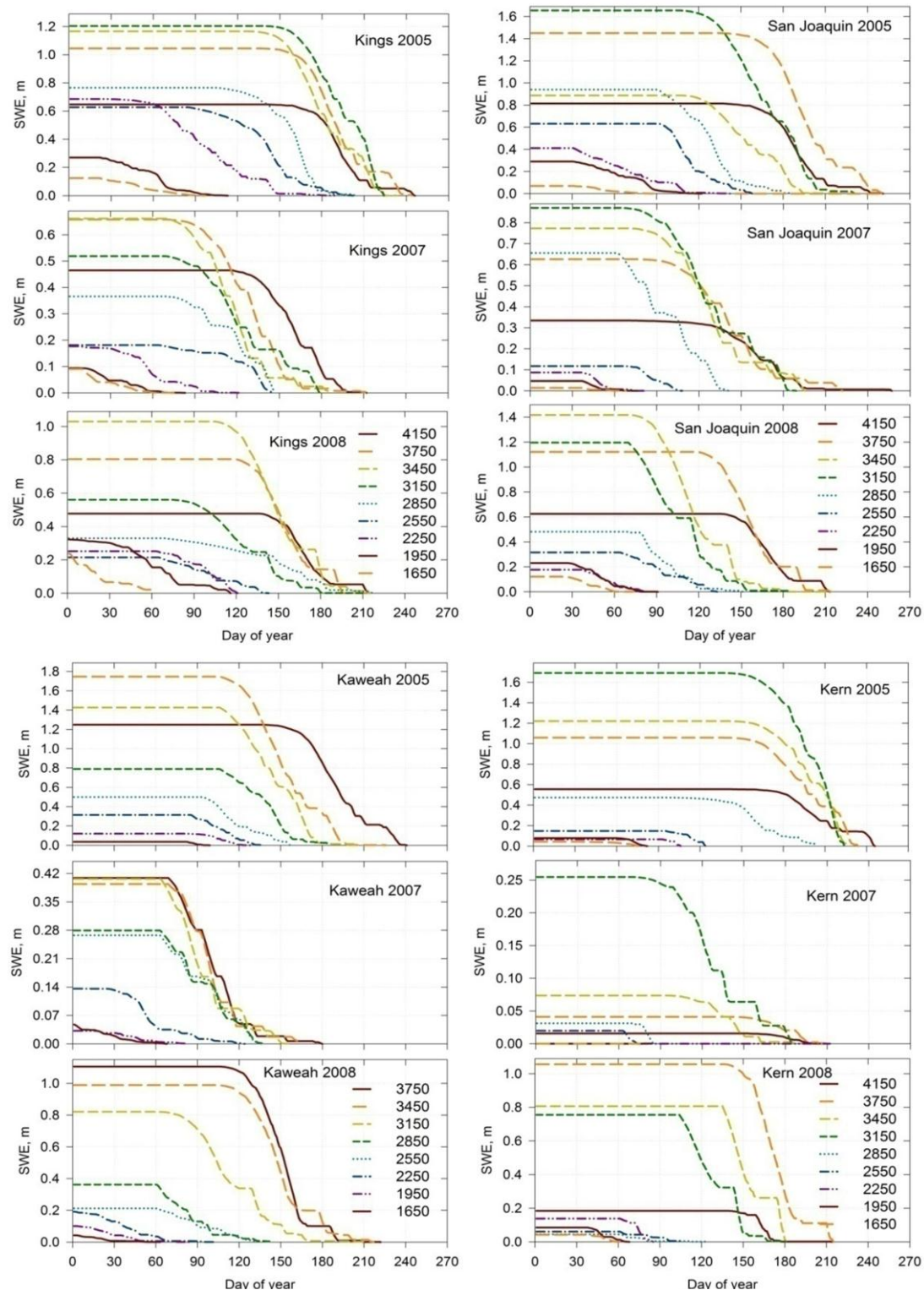
Back-calculated, snowmelt estimates from the depletion method are shown for the four basins for 2005, 2007, and 2008 on Figure 18. Recall that in this analysis daily snowmelt is summed back in time beginning with the day that SCA reaches zero. The summation is done back the day that snowmelt is zero owing to temperatures constantly below zero. This cumulative snowmelt is expressed as SWE. Because precipitation was small during snowmelt the product is indicative of SWE on the ground for that period. Note the steep gradient in SWE with elevation, with mainly elevations at 2400 m and above for the San Joaquin and Kings and above 2700 m for the Kaweah and Kern having the most snowmelt.

Results are qualitatively similar for the four basins, though snowmelt in the lower elevations shows up more in the San Joaquin and Kings River basins. Elevation gradients of the back-calculated seasonal snowmelt are illustrated on Figure 19. Note that seasonal snowmelt increases a similar amount with elevation in all basins, despite the approximately 20-50% more snowmelt in the San Joaquin versus the Kings, Kaweah, and Kern. The drop-off at the highest elevation in the Kaweah is due in part to the very small fraction of the basin in that elevation band and thus limited range of physiographic variability in that fraction.

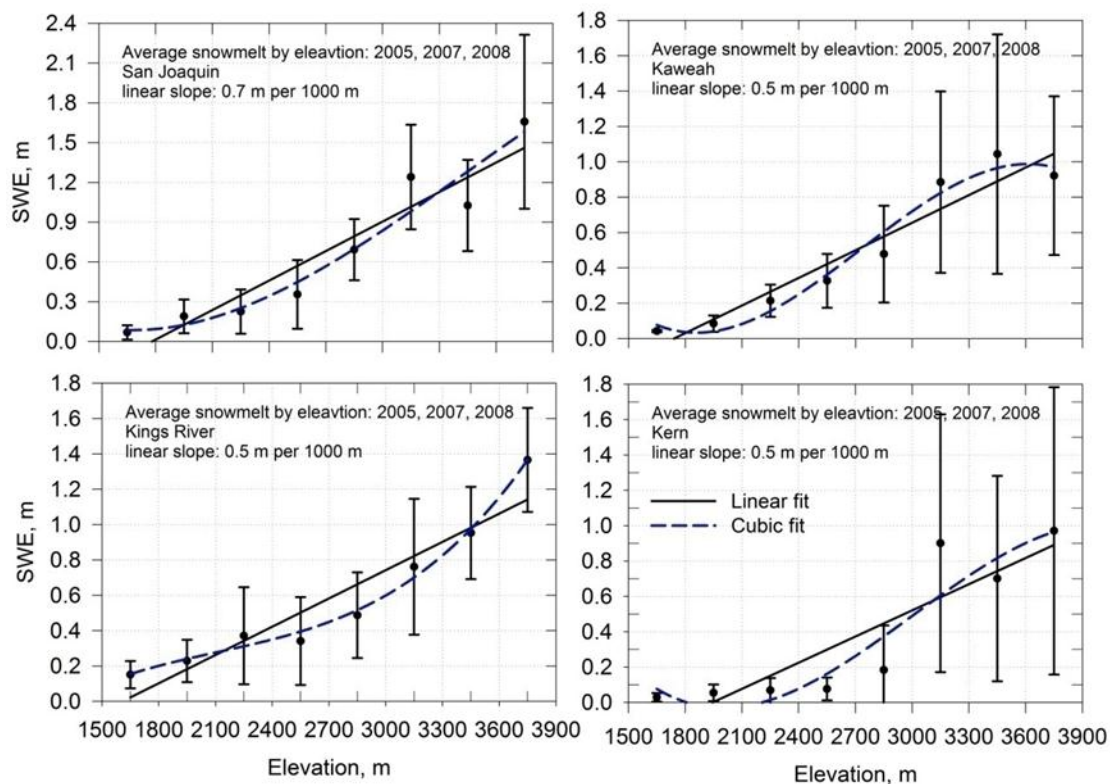


**Figure 17.** Elevation ranges of the San Joaquin, Kings, Kaweah, and Kern River basins.

SWE measured at snow pillows provides a short-term data set with the first snow-pillow measurements being archived in 1985, and significant snow-pillow data becoming available beginning in 1989. Thus a trend analysis of the available snow-pillow data cannot provide long-term reference conditions. It should be noted that trend analysis has been conducted using snow courses, manual monthly measurements of SWE, which provide the only long-term data on SWE (1930-present). Recent work has shown that the trend in SWE across the four basins has not significantly changed, and has even increased, even though the Western United States has experienced declines (Mote et al., 2005). A time series (2000-2009) of each of the snow pillow across the basins are provided to show the measured SWE in reference to the historical April 1 SWE as measured at the co-located snow courses (Figures 7a-7d).



**Figure 18.** Snowmelt by elevation band based on snow-depletion calculations using degree-day and satellite SCA. For the San Joaquin, Kings, Kaweah and Kern basins. Given by elevation band (legend in m).



**Figure 19.** San Joaquin, Kings, Kaweah, and Kern River basin SWE contributing snowmelt by elevation band based on depletion calculations showing three-year means, standard deviations and best fit to means. Data from Figure 17.

## Precipitation

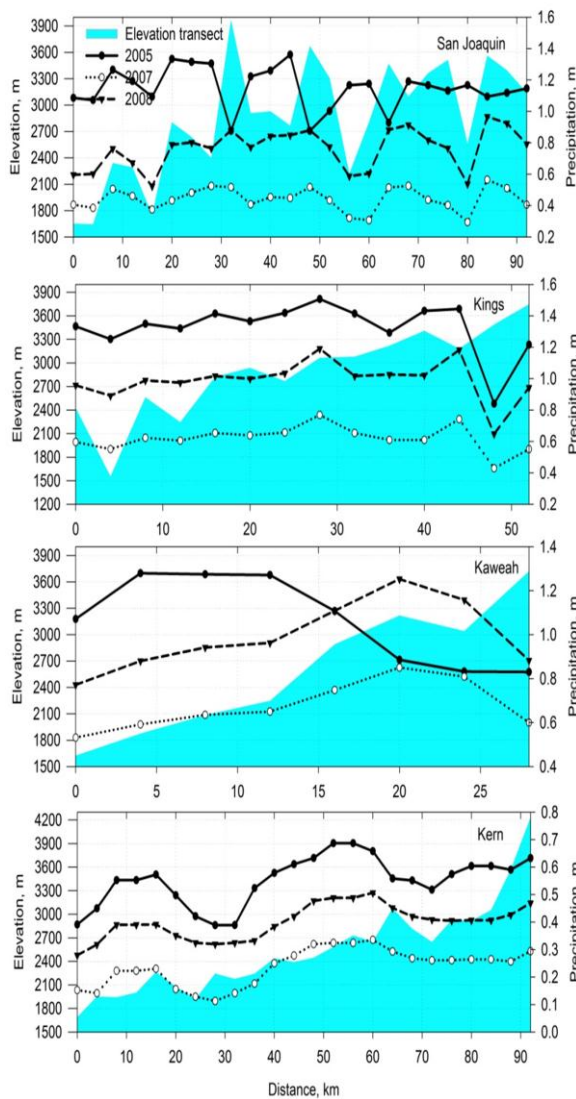
The annual, basin average PRISM 4-km gridded precipitation for the three years evaluated shows a weak relationship with elevation, and for 2005 in the San Joaquin and Kaweah there were declines in precipitation with elevation for both the October-March and annual precipitation (Figure 20a and 20b). This inconsistency is likely the consequence of the coarse spatial resolution across complex topography. The 800-m normal (1971-2000) gridded product does show a consistent increase with elevation, e.g. about 1.2 cm per 1000 m along a transect in the San Joaquin running from 1459 m to 3808 m, 1.8 cm per 1000 m along a transect in the Kings from 1788 m to 3933 m, 2.0 cm per 1000 m along a transect in the Kaweah from 1504 m to 3605 m, and 1.5 cm per 1000 m along a transect in the Kern from 1578 m to 4230 m (Figure 20c).

PRISM 4-km data do give an indication of annual and monthly precipitation, and there is a statistically significant correlation between these data and observations reported by CDEC and RAWS (data from WRCC) stations across the four basins (Figure 21). It should be noted that PRISM precipitation values are generally higher than those for RAWS stations. The precipitation gauge used at RAWS sites tends to underestimate winter precipitation. The error between the observed and PRISM annual values showed that 50% of the observed points had a percent bias of 17% or less (Figure 22) for the 2000-2009 period. Only 10% of the PRISM values had a percent bias exceeding 40% of the measured values, though 40% has a 20% or greater bias. These levels of bias point out the significant uncertainty that will be embedded in

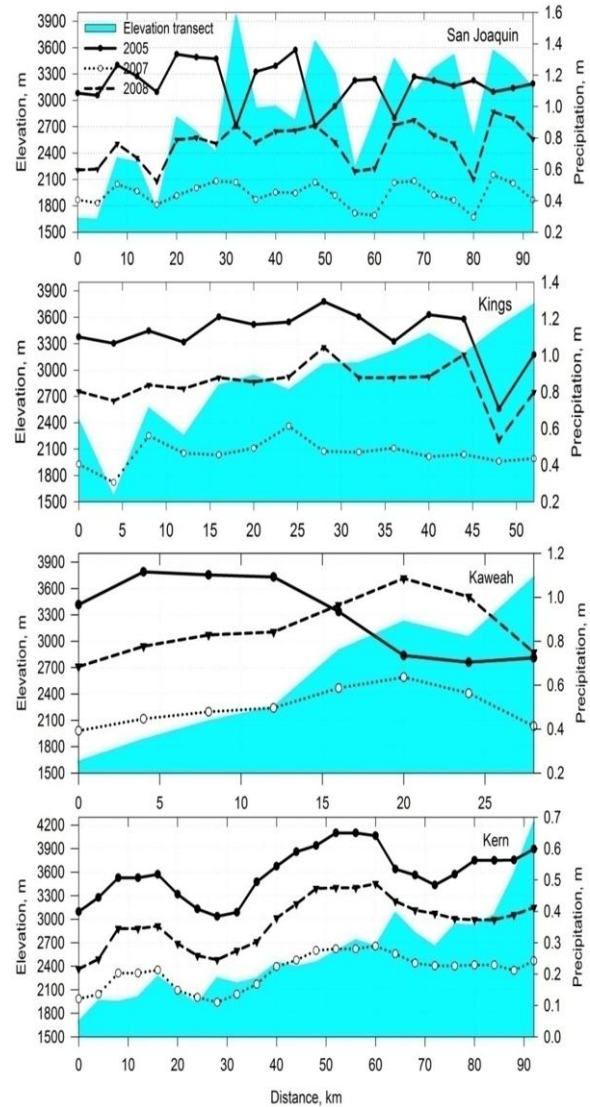


any water-cycle study, evapotranspiration estimates, or climate response studies using the available precipitation data.

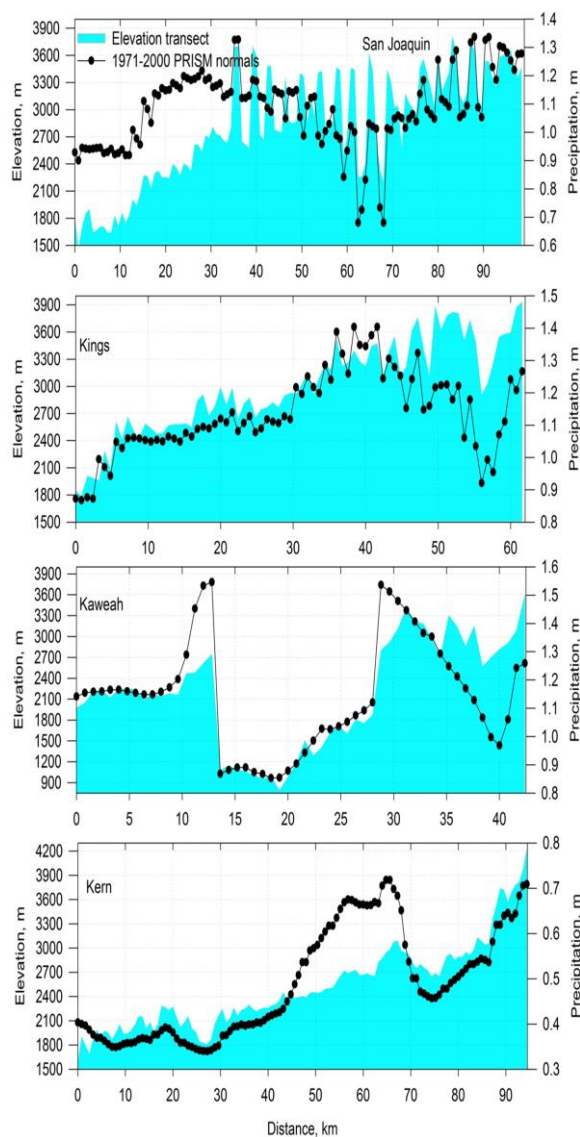
While the standard nationwide PRISM product, from which these values were taken, provides an estimate of patterns across the watershed that is needed for the hydrologic modeling, consideration should be given to improvement of this product using quality-controlled local data. To address this concern, corrections should be applied to PRISM across watersheds using station data when used as input for hydrologic modeling.



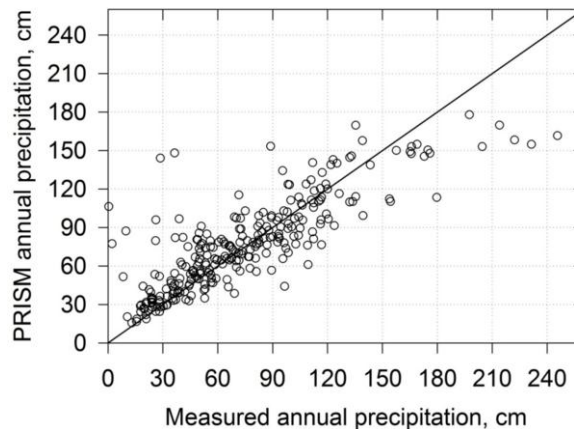
**Figure 20a.** PRISM 4-km Oct-Mar precipitation for Water Years 2005, 2007, 2008 extracted for single grid cell along elevation transects for the San Joaquin, Kings, Kaweah, and Kern River basins.



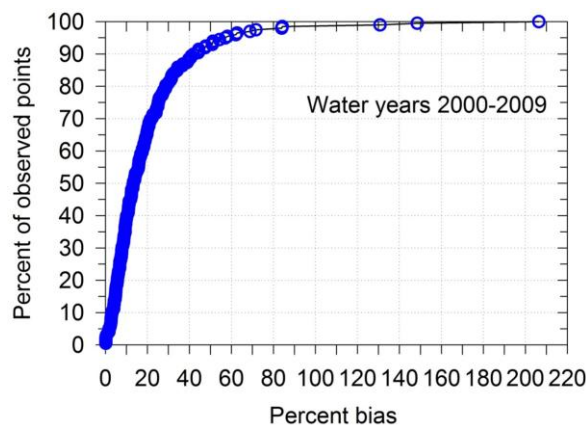
**Figure 20b.** PRISM 4-km annual precipitation for Water Years 2005, 2007, 2008 extracted for single grid cell along elevation transects for the San Joaquin, Kings, Kaweah, and Kern River basins.



**Figure 20c.** PRISM 800-m annual precipitation for Water Years 2005, 2007, 2008 extracted for single grid cell along elevation transects for the San Joaquin, Kings, Kaweah, and Kern River basins.



**Figure 21.** Comparison of PRISM annual precipitation (Water 2000-2009) estimates for single grid cells with values reported by CDEC and RAWS (WRCC) (observed) for stations within that grid cell for San Joaquin, Kings, Kaweah, and Kern River basins from stations shown on Figure 1.



**Figure 22.** Percent bias in the interpolated PRISM 4-km gridded precipitation product for 2000-2009 compared to the observed precipitation. Percent bias is the absolute value of measured minus PRISM divided by measured, expressed as percent.

As a further comparison, snowmelt calculated by the depletion method was compared to 2005, 2007, and 2008 October to March PRISM precipitation along the same transects as described above in the San Joaquin, Kings, Kaweah, and Kern River basins (Figure 23). Many PRISM precipitation estimates were higher than the estimated snowmelt especially at the lower elevations, since PRISM showed little or no elevation change across the elevation gradient, as discussed above. However, at higher elevations along the transect, total estimated snowmelt from the depletion method and the PRISM October-March precipitation had comparable results, as well as, in a few instances the depletion method estimated 25 to 50% more precipitation (maximum SWE) as shown in the Kings and Kaweah River basins. Again, these differences,

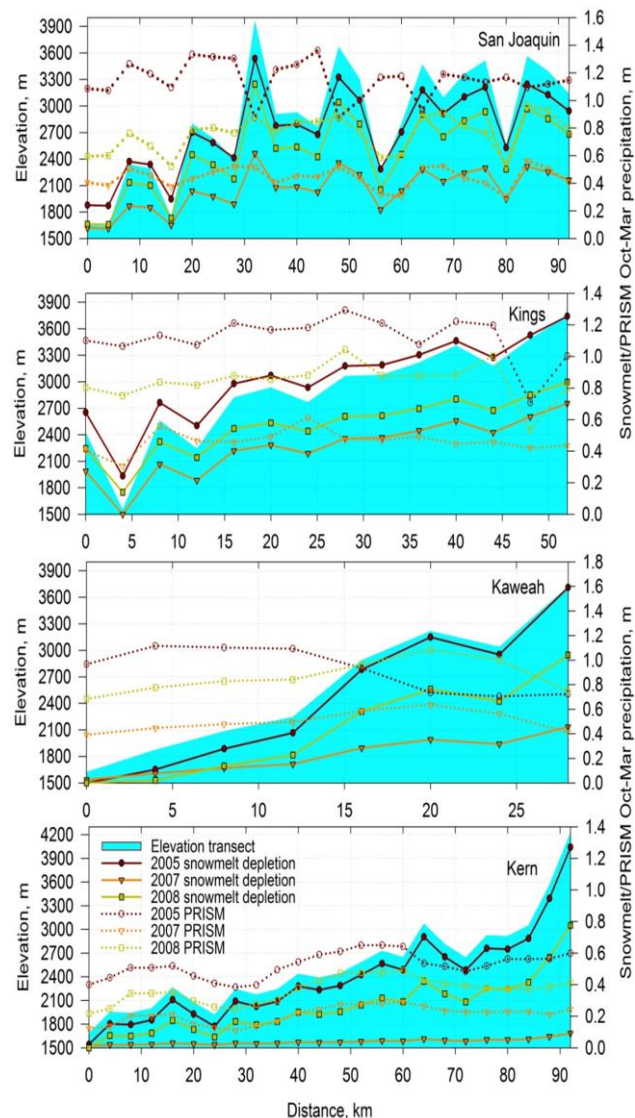


especially at the lower elevations, can be attributed to the coarse resolution of the PRISM 4 km product. There may also be rain effects at the lower elevations, as the snowmelt values do not include rain; PRISM is total precipitation. These analyses should be repeated as improvements to PRISM data become available.

## Paleoclimate

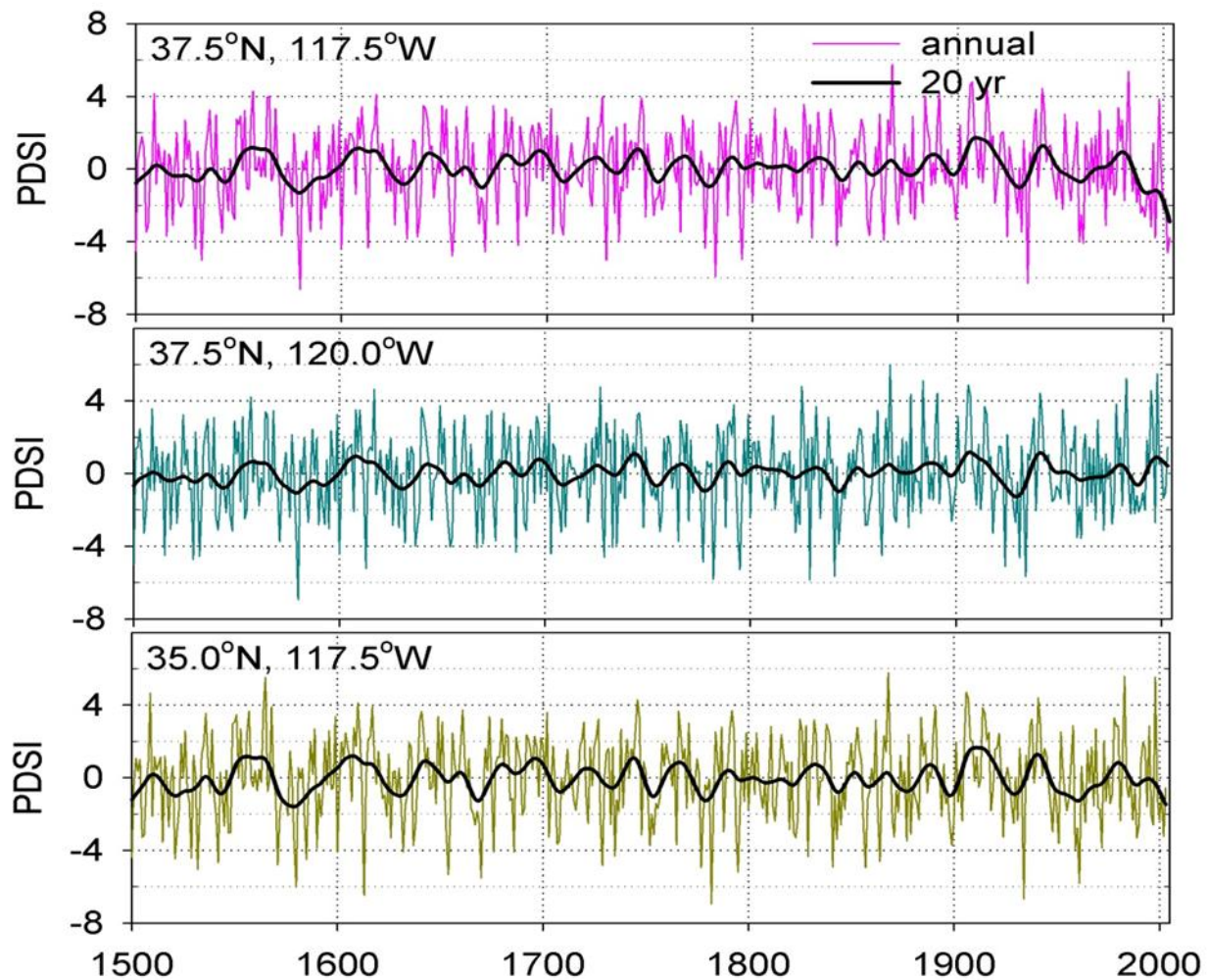
The PDSI reconstructions for the grid points that include the Parks extend back over 2000 years; the most recent 500 years are shown on Figure 24. The record for the two grids at 117.5°W, which includes parts of the Sierra Nevada but also Great Basin, clearly reflect the widespread drought conditions of the recent decade, the Southwest drought of the early 1950s, the extreme drought of the late 1500's, and other multi-year dry periods. Some of these events are less distinct for the next grid point west, which covers much of the high-elevation central Sierra Nevada, but also extends west in the San Joaquin Valley. Most notable is the lack of a sustained drought in the 1990s and early 21<sup>st</sup> century. Dry conditions in the 1930's and 1950's are apparent in both, with the latter more severe further south. The main message from these records is that decade-long dry periods are the rule rather than the exception, and that these dry periods tend to be relatively widespread, reflecting their link to larger, synoptic scale atmospheric circulation established by ocean temperature and pressure patterns.

Low-frequency patterns in the PDSI record track other climate indices, and the Sierra Nevada tends to respond to common influences. Mono Lake sediments have recorded five major oscillations in the hydrologic balance of the region in the period 1700-1941. These oscillations have been correlated with tree-ring-based oscillations in Sierra Nevada snowpack, and reconstructions of the Pacific Decadal Oscillation (PDO) index, indicating that major oscillations in the hydrologic balance of the Sierra Nevada correspond to changes in the sign of the PDO (Benson et al., 2003). Extreme droughts occurred during PDO maxima, at approximately 60 to



**Figure 23.** Estimated snowmelt using the depletion method and Oct-Mar precipitation from PRISM along an elevation transect for 2005 (wet), 2007 (dry), and 2008 (average) extracted for single grid cells along elevation transects for the San Joaquin, Kings, Kaweah, and Kern River basins.

80 year reoccurrence intervals (1710, 1770, 1850, and 1930). The PDO for the instrumental period shows an oscillation over the past 10-15 years between negative and positive.



**Figure 24.** PDSI values for 3 grid points that include parts of the Southern Sierra Nevada. Grid 059 is centered on the California-Nevada border just north of the Inyo-Mono county line, grid 047 is centered on the town of Mariposa, west of Yosemite NP, and grid 060 is centered in the desert of southeastern California.

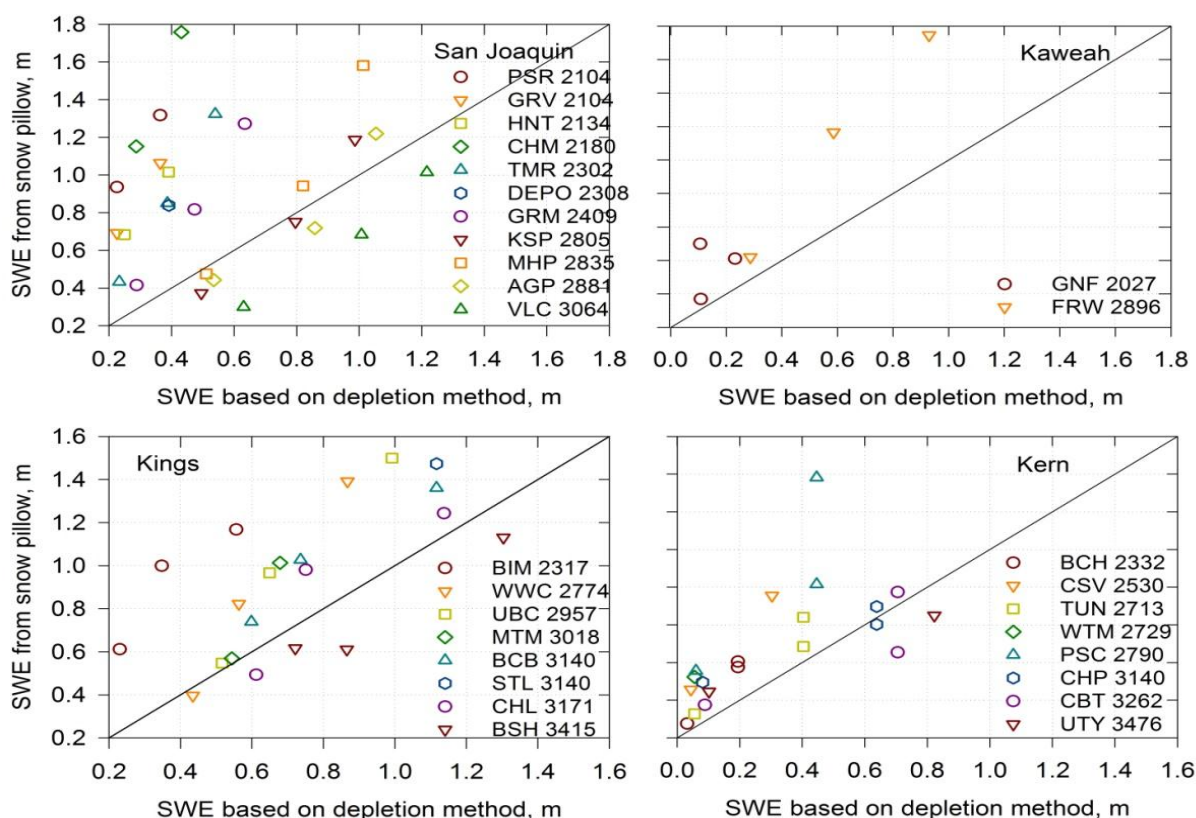
Another locally relevant paleoclimate reconstruction comes from the Bristlecone Pine tree ring data, largely in the White Mountains (Hughes and Graumlich, 1996). The sort of oscillations noted above for the millennial scale are even more apparent in the 8-millennium record of reconstructed precipitation for the White Mountains.

Several lines of evidence point to precipitation in the Sierra Nevada being influenced by processes that affect west slope precipitation and runoff, as well as that in the Great Basin. For example, Graham and Hughes (2007) reported a correlation between reconstructed streamflow on the western slope of the Sierra Nevada (Merced River) and Mono Lake levels. With this in mind, it is instructive to examine the chronology developed for northern Sierra Nevada streamflow (Meko, 2001 and later reports). Again, this long-term record highlights decadal and

longer droughts. Much of Western North America has recently been in a major multi-year drought, of a scale and intensity that has occurred several times during the last 2000 years, especially before the middle of the second millennium AD (Hughes and Diaz, 2008). Were one of these naturally occurring multi-decadal droughts to recur, the consequences for water supply and other important ecosystem services would be extreme. A multi-decade interval with a 35% decrease in precipitation could be economically and environmentally devastating for resource management in the American West.

## Spatial and temporal analyses

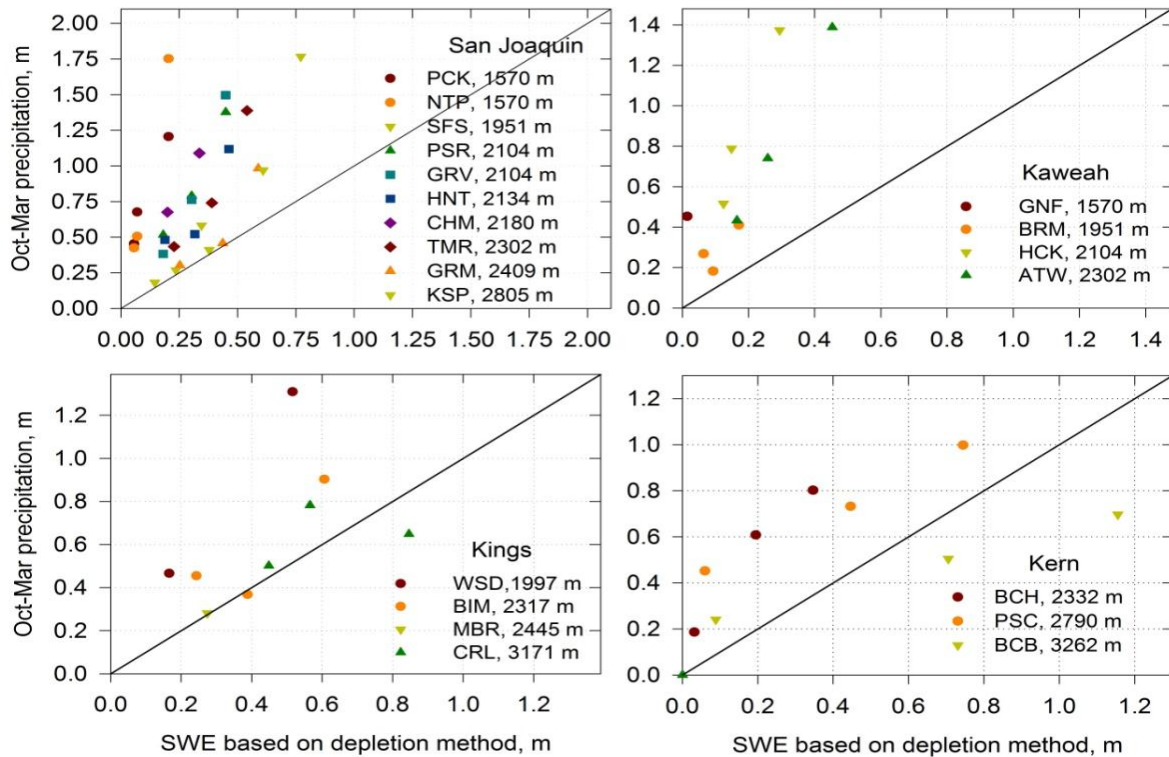
One possible comparison for the calculated SWE and melt rates are the observations made at snow pillows. It should be recognized that the snow pillows are not representative of the terrain in the basins, but should be in the same range as the band-average total snowmelt, or SWE estimated by the depletion method. SWE from the calculation by elevation band and the snow pillow peak SWE values are generally in the same range, but the snow pillows melt out much faster. That is, a significant amount of snowcover persists in the basins and contributes snowmelt well after the snow pillows are snow free. Comparing just the maximum seasonal SWE at snow pillows, and snowmelt by the depletion method interpolated from the 300-m elevation band analysis to the elevations of the snow pillows, illustrates that the pillow values are generally higher than that estimated from the degree-day and SCA calculation (Figure 25). This is expected since the calculated values are adjusted for fractional snowcover, and the snow pillows are fully snow covered until they melt out.



**Figure 25.** San Joaquin, Kings, Kaweah, and Kern maximum SWE by elevation band based on depletion calculations compared to peak SWE from the snow pillows in the basins for 2005, 2007, and 2008.

Another point of comparison for the back-calculated snowmelt was a comparison with precipitation measured at the mountain stations. Comparing October through March precipitation with back-calculated snowmelt for the same elevation shows precipitation gauge measurements to be higher (Figure 26). Again, it should be recognized the precipitation gages are point measurements in heterogeneous terrain, and cannot be assumed to represent spatial averages.





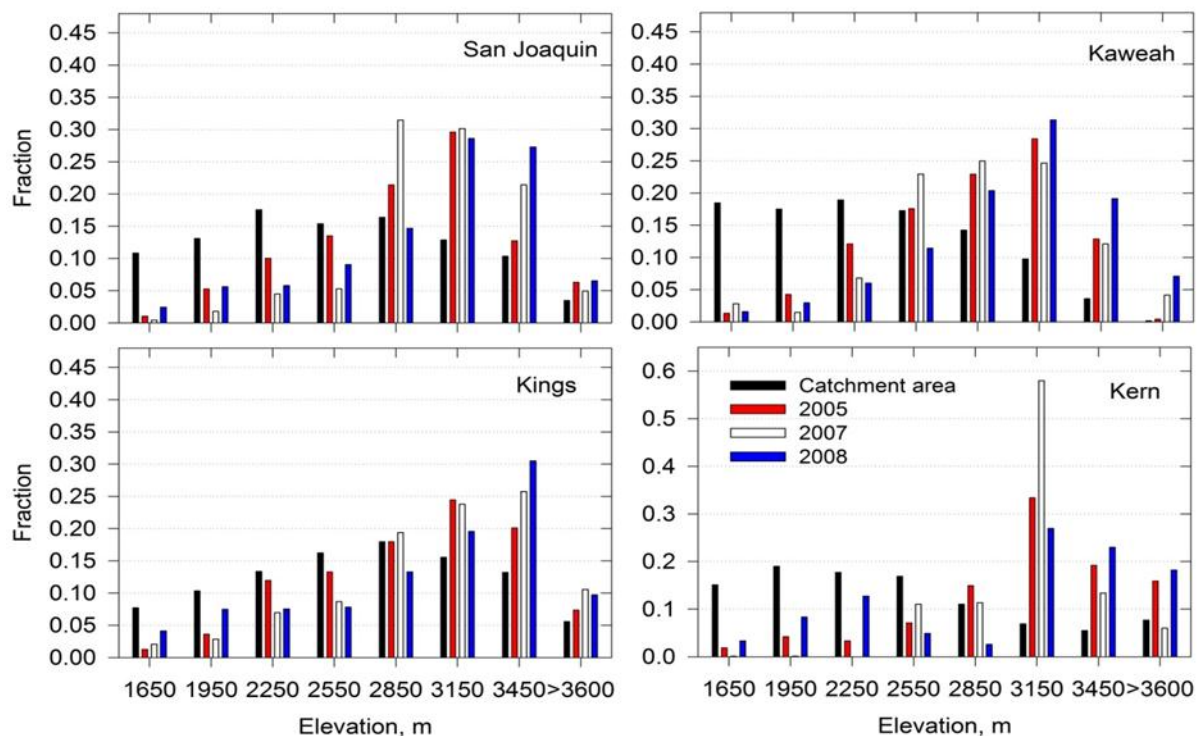
**Figure 26.** Maximum SWE by elevation band based on depletion method and reported precipitation from rain gages in the San Joaquin, Kings, Kaweah, and Kern River basin.

## Snowmelt

The majority of the snowmelt in the San Joaquin, Kings, Kaweah, and Kern was derived from elevations above 3000 m (Figure 27). In 2005 when the snowpack in San Joaquin, Kings, Kaweah, and Kern were 167%, 163%, 159%, and 184%, respectively, of the historical April 1 snow course average, 50-65% of the total snowmelt volume was derived from the area above 3000 m. The area above 3000 m accounts for 26%, 35%, 23%, and 20% of the total area in the San Joaquin, Kings, Kaweah, and Kern, respectively. The mid-elevations, 2400-3000 m, contributed similar amounts, 26-40%, of the total snowmelt volume. These mid-elevations were 32%, 34%, 41%, and 28% of the total area of the San Joaquin, Kings, Kaweah, and Kern, respectively. Below 2400 m only 17% of the total snowmelt was derived from these elevations in the San Joaquin, Kings, and Kaweah; while in the Kern only 7% of the total snowmelt volume was derived below 2400. The lower snowmelt volume in the Kern is likely a result of being the furthest South in latitude, and the South orientation of the basin, which is unique to the either a West or East orientation of river basins throughout the Sierra Nevada. In 2007, when San Joaquin, Kings, Kaweah, and Kern were 43%, 40%, 40%, and 21% of the historical April 1 average, trends were similar, but a smaller fraction of the total snowmelt volume was derived from the lowest elevations, while elevations above 3000 m accounted for much of the difference in snowmelt volume in the wetter (2005) versus drier (2007) year. In the San Joaquin, Kings, Kaweah, and Kern, 7%, 12%, 11%, and 1% of the total snowmelt volume were derived from below 2400 m. The mid-elevations, 2400-3000 m remained relatively constant with the San Joaquin and Kings, contributing 37% and 28% of the their respective snowmelt volumes, while the Kaweah increased to 48% of the total snowmelt volume, and the Kern declined to 22% of the



total snowmelt volume. Elevations above 3000 m contributed 56%, 60%, 41%, and 77% of the total snowpack volume in the San Joaquin, Kings, Kaweah, and Kern.



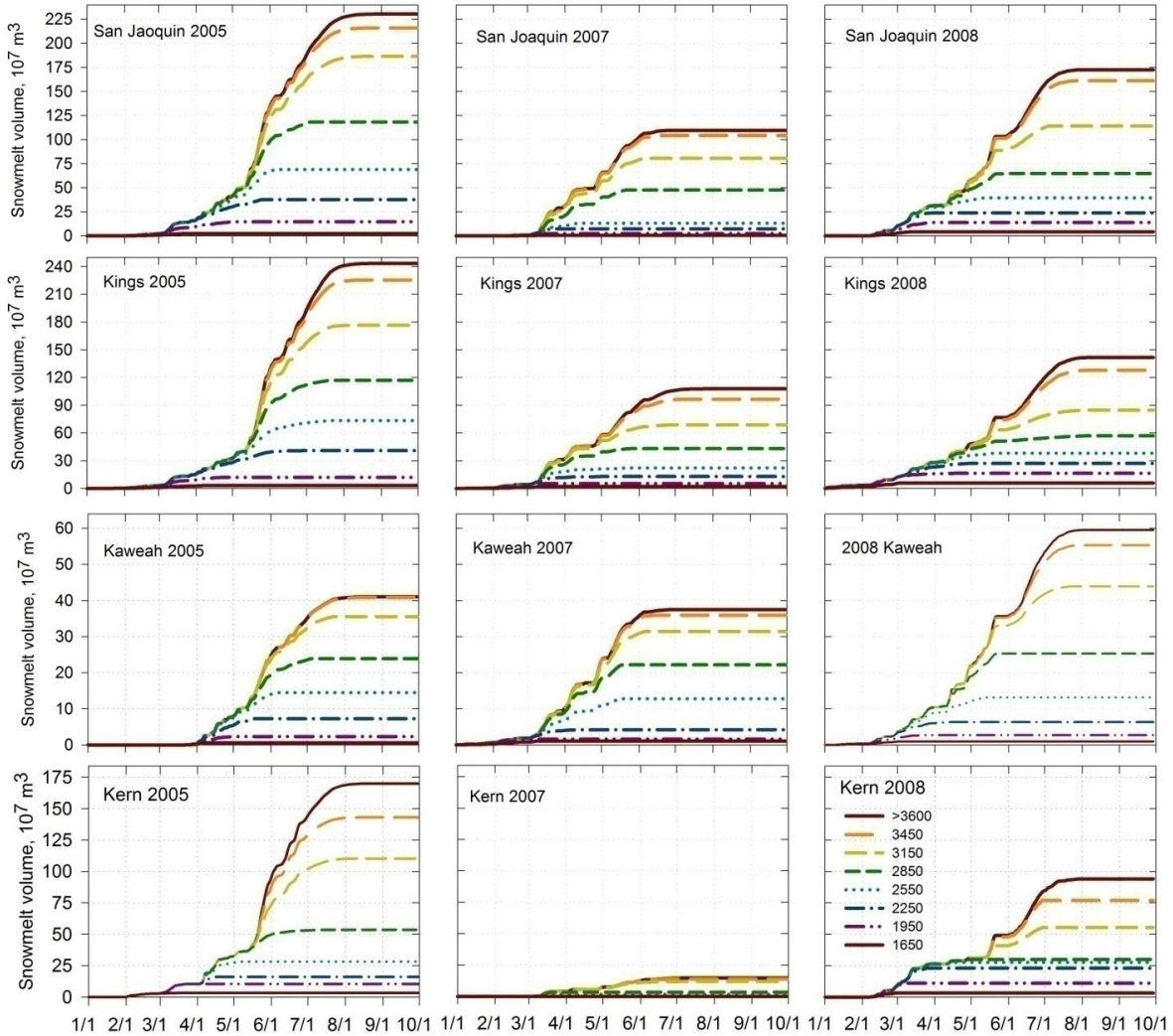
**Figure 27.** Fraction of annual snowmelt by elevation band for San Joaquin, Kings, Kaweah, and Kern River basins. The sum of fractions for all elevation bands for a given year is one. Average for 2005, 2007 and 2008.

Elevations below 2400 m contributed 100% of the March 2005 snowmelt in all four basins, declining to 30% (Kern) to 75% (Kings) in April and 0-10% May (Figure 28). Mid elevations (2400-3000 m) contributed no snowmelt in March, increasing to 26% (Kings) to 70% (Kern) in April, 36% (Kern) to 59% (Kaweah) in May, and dropping to 10% (Kern) to 40% (Kaweah) by June and nearly disappearing by August. Elevations above 3000 m contributed no snowmelt to the monthly total in March, increasing to 30% (Kaweah) to 39% (Kings) in May, and 89% (Kaweah) to 100% (Kern) in July.

In 2007 the snowmelt pattern was similar, but the relative snowmelt contributions were one month earlier due to the smaller snowpack, which was 43% of the historical snow course averages. In the San Joaquin, Kings and Kaweah basins the lower elevations (<2400 m) contributed 100% of the snowmelt during February declining to near zero by April 30. In the Kern contributions from the lower elevations were insignificant through all months. The mid elevations were contributing 45-60% of the March snowmelt, increasing to 50-65% for April, and declining to zero in June. Elevations above 3000 m contributed 15-35% of the March snowmelt increasing to 100% for June, with a steady decline to zero in July.

In 2008, which is considered an average year, based on the MODIS period of record analyzed here (2000-2009), the snowmelt patterns were similar to 2005 and 2007. The lower elevations below 2400 m began contributing to snowmelt in February, with no contributions from

elevations above 2400 m. By April 1 all elevations were contributing to snowmelt, from 40% (Kaweah) to 89% (Kern) below 2400 and 19% (Kern) to 59% (Kaweah) for elevations 2400-3000 m. In April snowmelt contributions shifted, with 55-75 of the snowmelt deriving from the mid elevations and 25-35 from above 3000 m, with elevations below 2400 m declining to zero by July, and the upper elevations declining to zero in August.

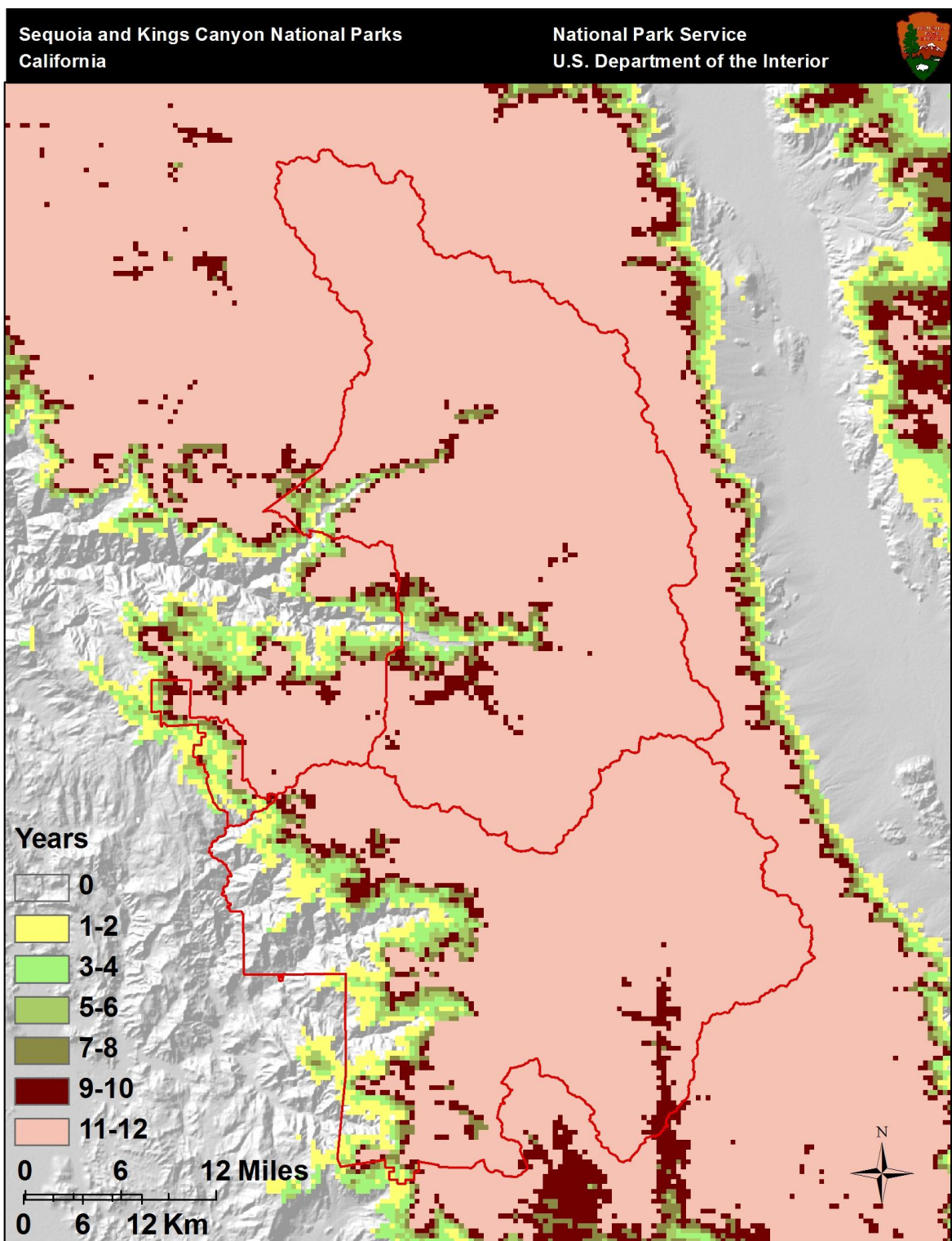


**Figure 28.** Cumulative snowmelt by basin for San Joaquin, Kings, Kaweah, and Kern River basins. by elevation band (legend in m).

The average lapse rate corresponds to approximately a 1.7-2°C change in average temperature for each 300-m elevation change. This suggests that a long-term change in average temperature of 2°C corresponds to an elevation change of about 300 m. Substituting space for time, current snowpack conditions at some elevation can be expected to indicate snowpack conditions at an elevation that is 300 m higher under a 2°C warmer climate; e.g., conditions at 2400 m today are what conditions would be like at 2700 m under a 2°C warmer climate, or 3000 m under a 4°C warmer climate.

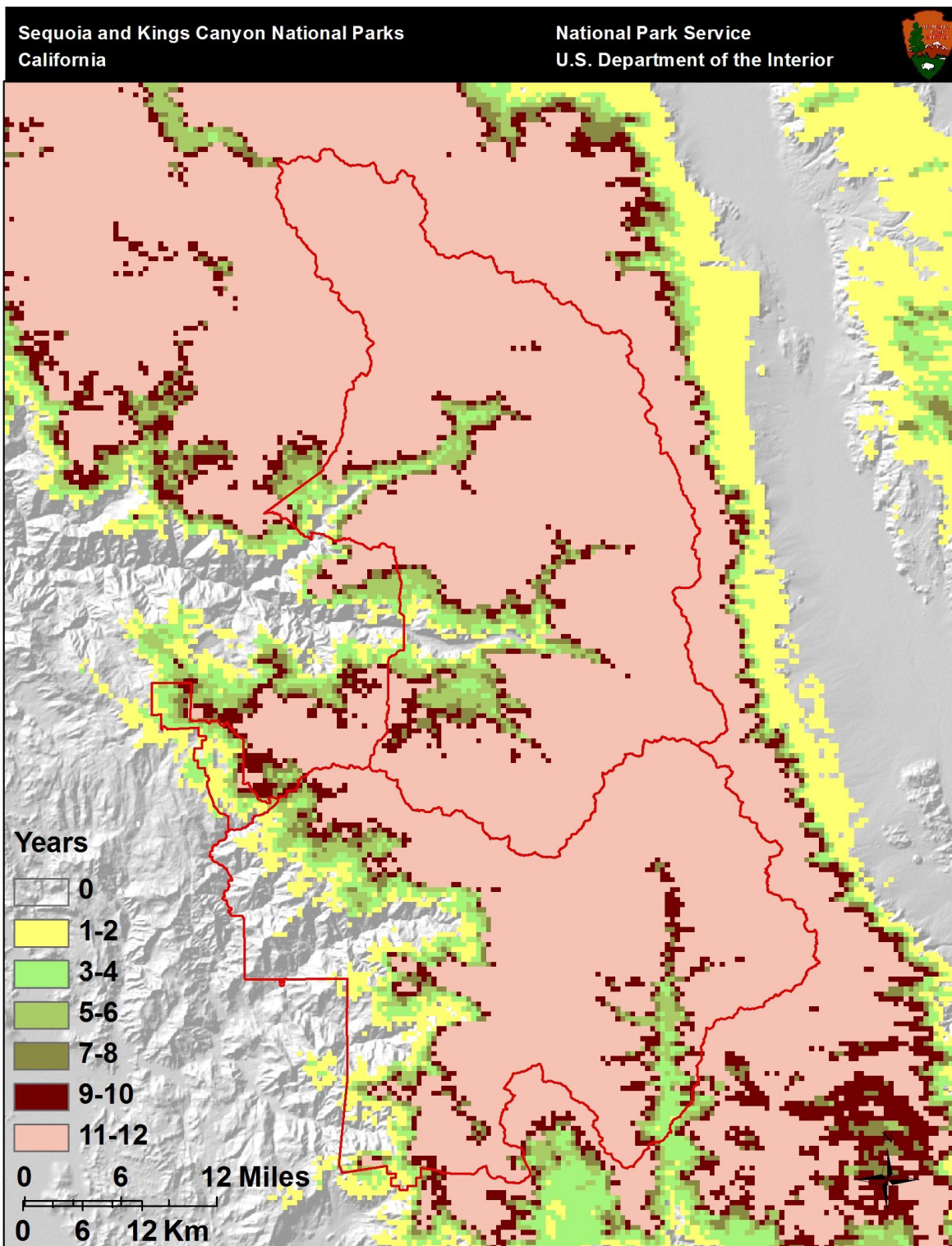
### **Snowcover Persistence**

Areas with persistent snowcover are potentially less vulnerable to small temperature changes than areas that have snowcover only in some years but not others. Figures 29a-d show areas that for the period 2000-2011 had more versus less persistent snowcover for the first of each month, March through June. Note that much of the Parks consistently have some snow March and April 1, with the middle and lower elevations of the Kngs and Kaweah basins having only intermittent snowcover on those dates. These patterns are accentuated in May and June, with only the highest elevations retaining snowcover.

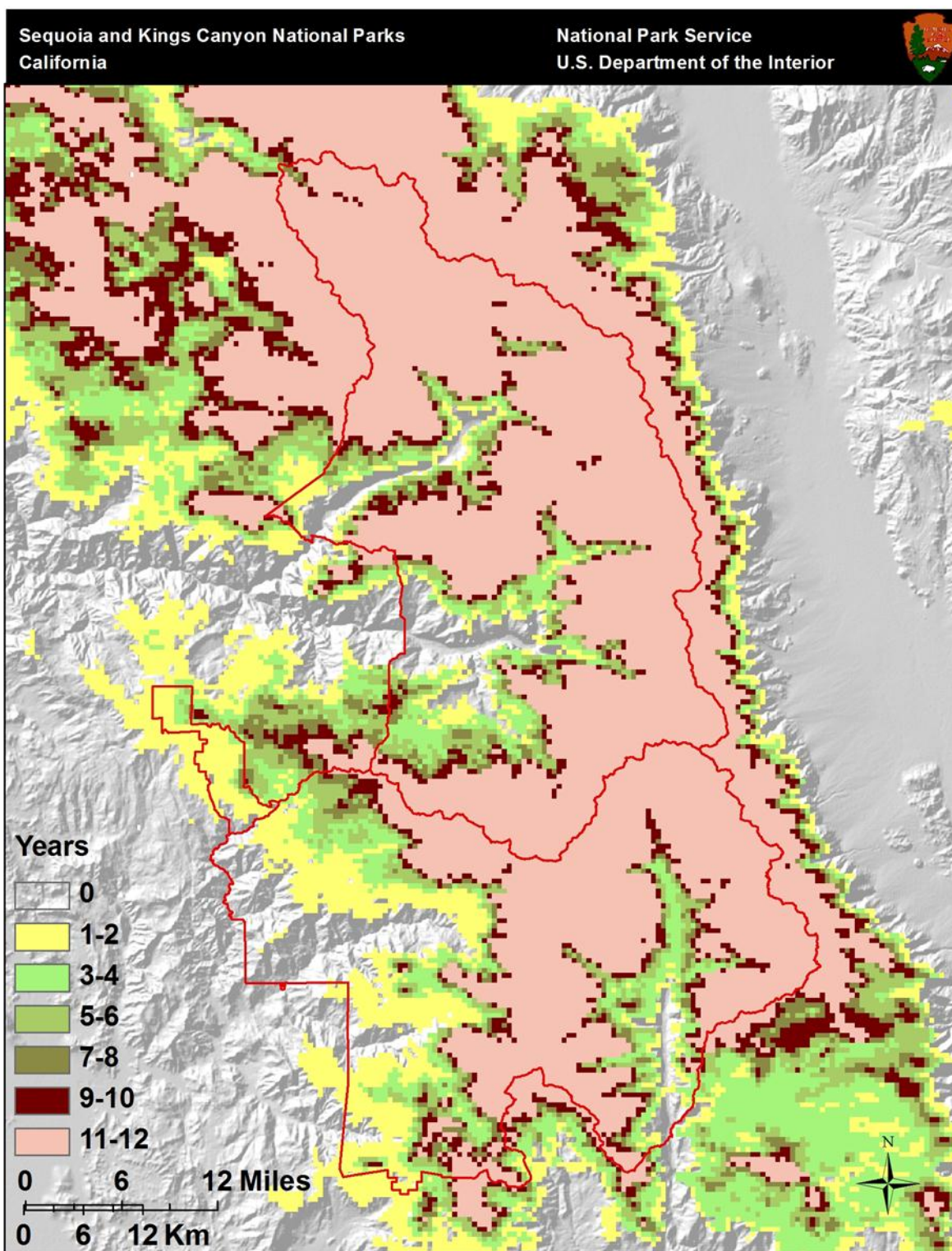


**Figure 29a.** Snowcover persistence, or number of years each pixel had >20% snowcover on March 1.



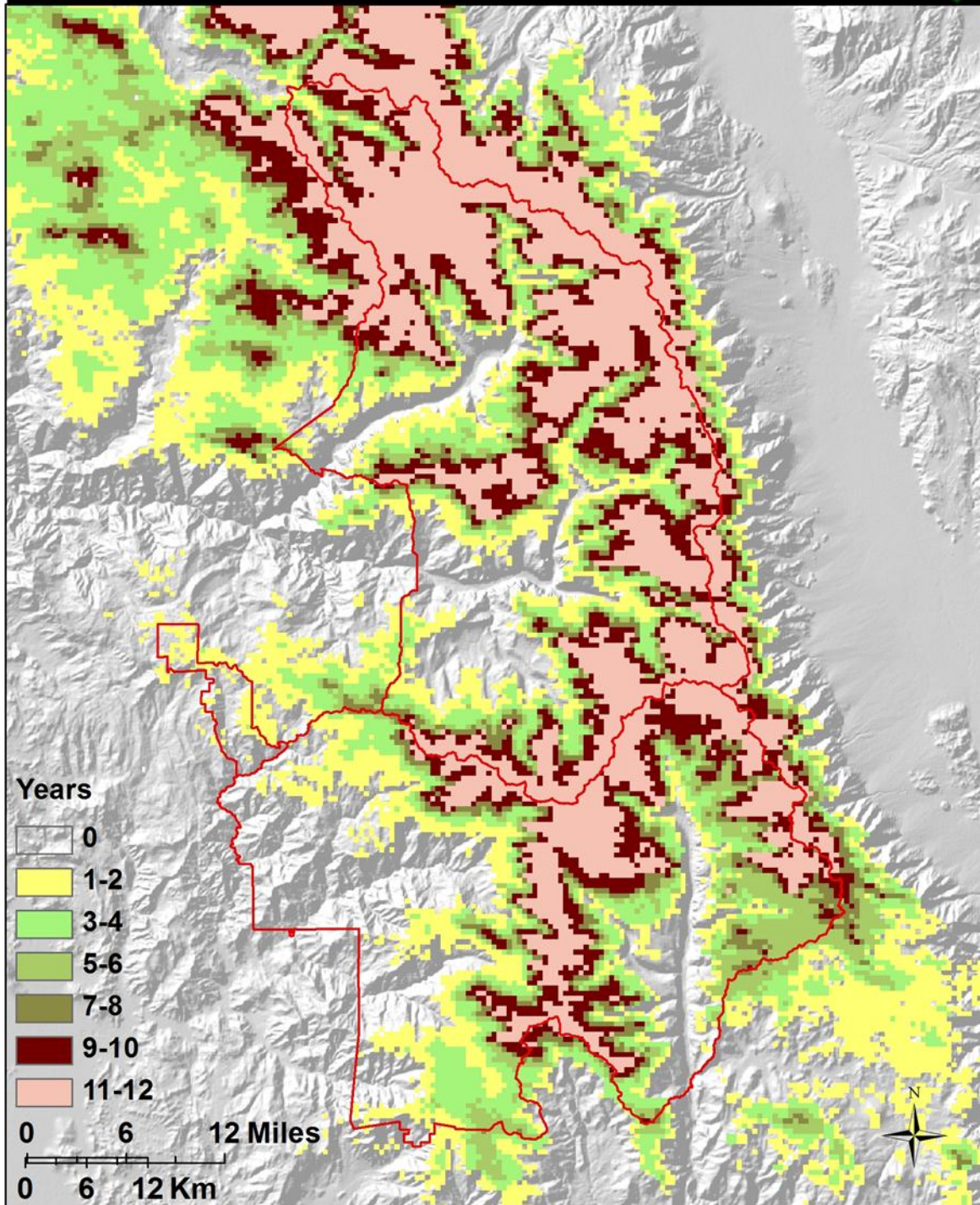


**Figure 29b.** Snowcover persistence, or number of years each pixel had >20% snowcover on April 1.



**Figure 29c.** Snowcover persistence, or number of years each pixel had >20% snowcover on May 1.





**Figure 29d.** Snowcover persistence, or number of years each pixel had >20% snowcover on June 1.





## Analysis Uncertainty

Broad temperature patterns are known with some confidence across the Parks, as indicated on Figure 11. That figure shows that most points fall close to a regression line with elevation. Another view of the variability is to index all temperatures to a common elevation (Figure 30), in this case 3000 m. Any elevation could be used, with the same variability in temperature. The 1-3 °C standard deviation reflects the variability in temperature due to physiographic variability at a common elevation, and for an average 6.8°C ground-based lapse rate is equivalent to about 150-450 m elevation. Cold-air drainage at night may be partly responsible. Comparing Lodgepole and Grant Grove on Figure 4, it is apparent that the two sites, which are at comparable elevations, have a similar daily maximum temperature but a daily minimum that differs by about 3°C. The implication for snow resources is that there is already about a 2-4 week difference in melting out of snow within an elevation band, comparable to the difference between melting across adjacent 300-m elevation bands.

Some of the uncertainty may also be in the temperature records themselves. Several manual corrections were made to the data, but it is apparent that several records receive little calibration and quality control over the years, and further analysis will be needed to examine sub-daily temperatures in any detail.

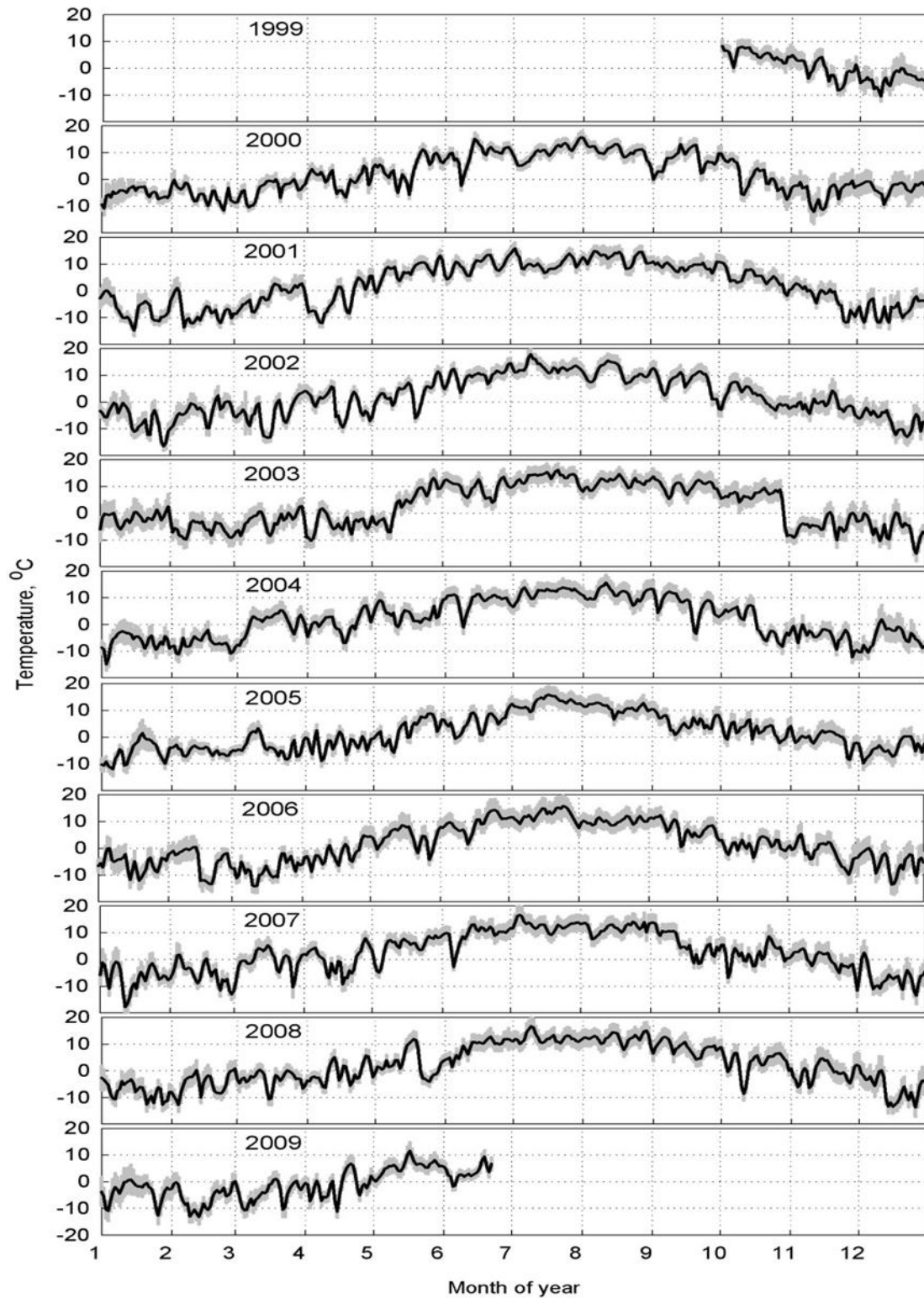
There is also uncertainty in the calculated snowmelt (Figure 18), whether by pixel or by elevation band. The current analysis used a temperature-index approach, calibrated to data within the watershed, and is appropriate for the data available. It is judged to be a better estimate of snowpack than that developed from climate models, which are generally indexed with the long-term PRISM estimates. Note that PRISM estimates of precipitation at elevations dominated by snow are generally higher than the estimates from the back calculation of snowmelt for the four Westside southern SIEN basins evaluated. Further analysis of this bias is warranted. Improvements in the snowmelt estimates should come from refinement of the snowmelt energy balance, which should be based on improved vegetation data for the basins of interest.

## Interactions with other focal resources

Snowpack, precipitation and temperature controls are central to stresses on multiple focal resources. As indicated on Figure 2, precipitation and snowpack, together with soil-water storage, determine the amount of water available for ET and streamflow. Snowpack is a critical SEKI resource that provides seasonal storage of water for soil moisture, lakes and streams.

## Stressors

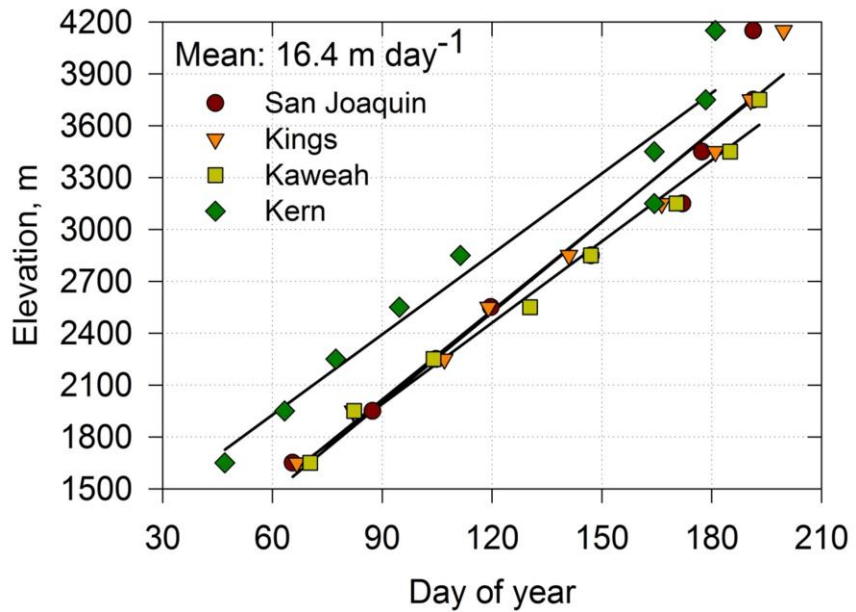
Three climate change scenarios are considered, 2, 4 and 6°C. Note that each 2°C rise in temperature corresponds to the current temperature patterns about 300 m lower in elevation. Note that each successively higher 300-m elevation band melts out about 20 days later than the elevation band below it, in warm and cool springs, deep or shallow snow, high or low elevations (Figure 31). Thus to a first approximation, this rate should continue in a warmer climate. However, melting will be earlier.



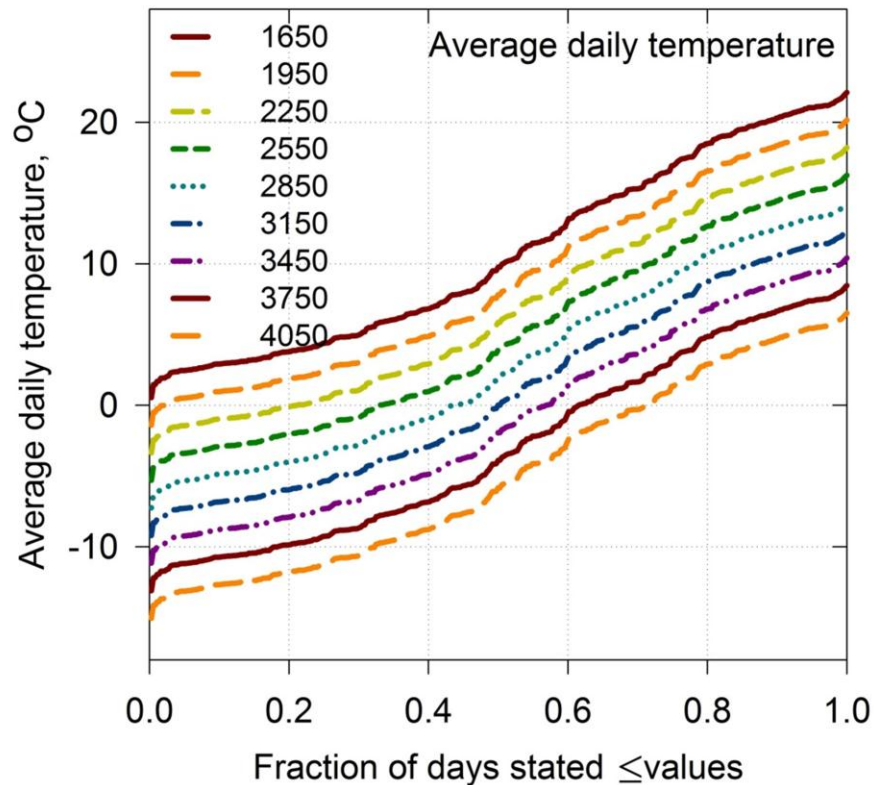
**Figure 30.** Daily temperature from the stations on Figure 4 indexed to 3000 m elevation using 6.8oC ground-based lapse rate. Mean and standard deviation of all 51 stations shown.

The current distribution of temperatures by elevation band shows that at elevation 3000-3300 m about half of the year has a daily average temperature below zero (Figure 32). At 2100-2400 m, about 20% of the year has a temperature below zero. With temperature increases of 2, 4 and 6°C, at 3000-3300 m elevation only 43, 33 and 20% respectively of days will have averagedaily temperatures below zero. In terms of number of days, at 3000-3300 m this represents a change from 180 to 150, 110 and 75 days, respectively (Figure 33).

In terms of snowmelt, the current calculation is based on degree days, allowing projecting future snowmelt based on temperature. Using variability in current temperatures for individual years for one elevation (3000 m), expressed as cumulative degree days (Figure 34), shows that in a warmer year (2007), a given value of cumulative degree-day



**Figure 31.** Snowmelt progression, or average increase in snowline during melt season. Values on this graph are indexed at 20% SCA rather than zero.

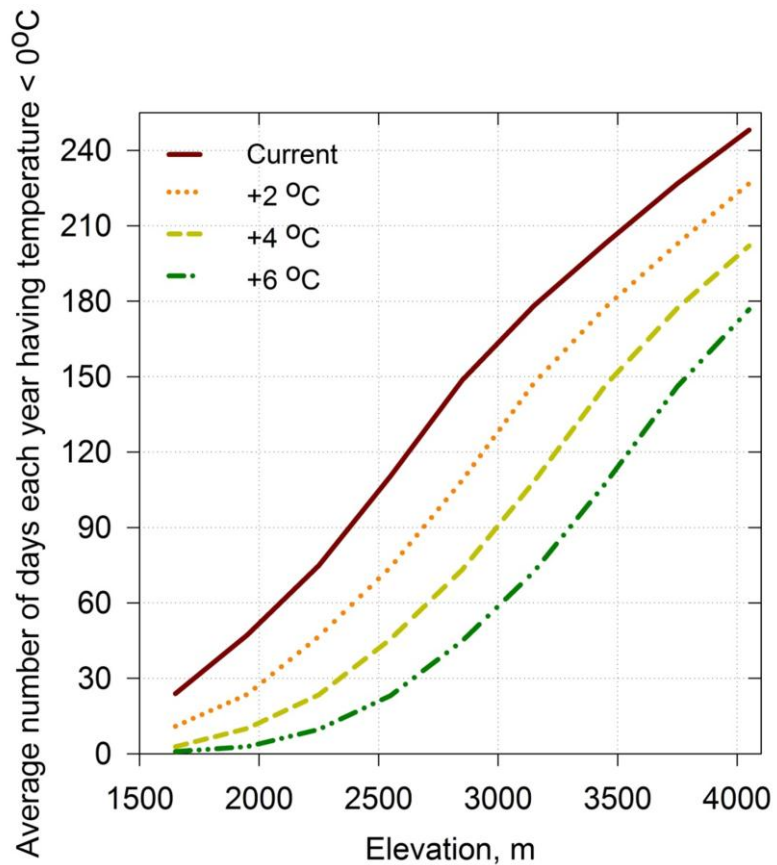


**Figure 32** Distribution of daily average temperature for each 300-m elevation band based on mean values from Figure 28, averaged over 2000-2009 (elevations in m)

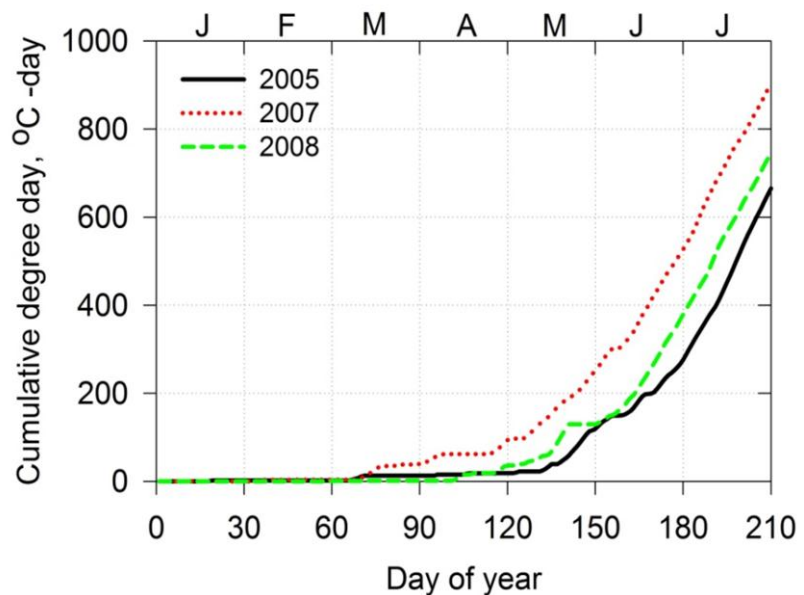
is reached about 30 days before the same level is reached in a colder year (2005). That is, the potential for snowmelt occurring by a given date varies nearly as much as is the expected change in mean of a 2oC warming. So even though both warm and cold months occur in most years, cumulative degree days clearly show warm and cold years for snowmelt.

This same 30 day or one-month advance in snowmelt per 2oC of warming noted above also shows up in examining the projected cumulative degree days for the 3000-m elevation (Figure 35). Other elevations show a similar pattern (data not shown). Cumulative degree days are an index of the cumulative potential for snow to melt. This 30-day change in the date that a given level of cumulative degree day is reached per 2oC change in temperature represents a real advance in snowmelt. As noted above, since the lapse rate is 6.8oC, each 2oC higher temperature change also represents a 300-m lower elevation under present temperatures.

In summary, several temperature calculations show that on average, a 2oC warming can be expected to shorten the snow season at a given elevation by an average of one month. This change is



**Figure 33.** Mean number of days temperature is below zero currently, and with a 2-6°C increase in average temperature.



**Figure 34.** Cumulative degree day for 2005, 2007, and 2008, based on Figure 30, for 3000-m elevation.



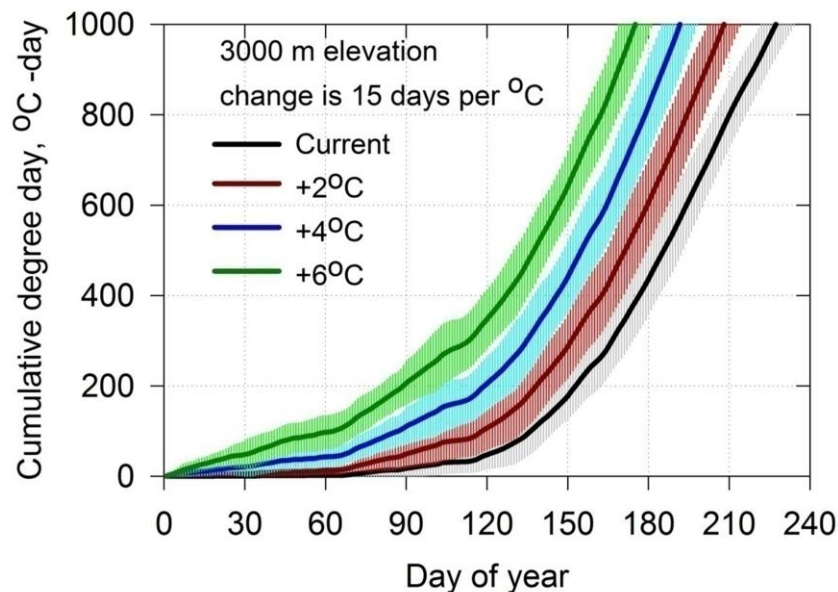
approximately the same as the current interannual variability in persistence of an equivalent amount of snow. The actual interannual variability in snow persistence is somewhat longer because both precipitation amount and temperature vary, i.e., snow will persist much longer in a cold, wet year than in a warm, dry year.



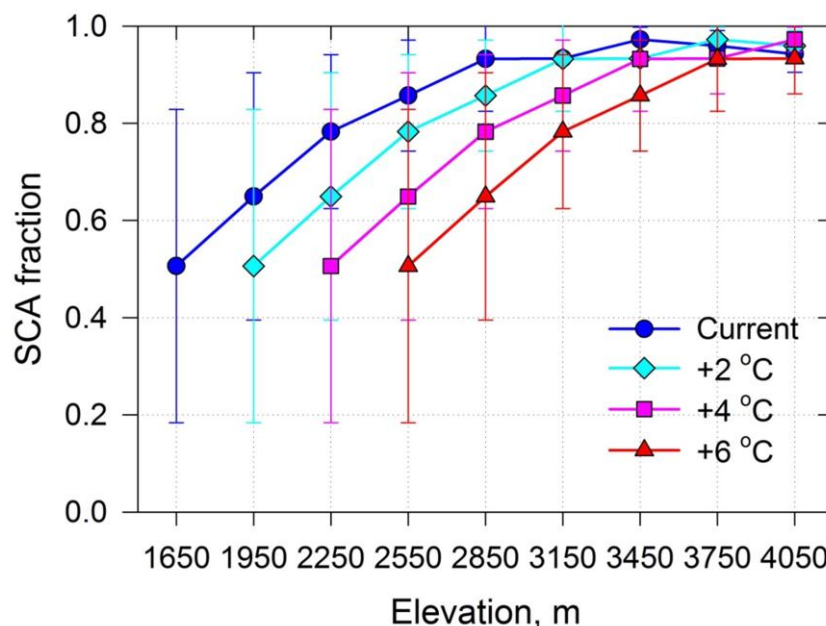
## Assessment

Temperature increases will shift the amount of snow covered-area as less precipitation falls as snow. To a first approximation, this can be viewed as a shift of 300 m in current SCA patterns as being equivalent to 2°C (Figure 36). Currently across all four basins there is an elevational dependence of winter SCA with each successive lower 300 m elevation band having greater inter-seasonal variability. That is, higher elevations, which are consistently below freezing much of the winter, have more consistent snowcover than do lower elevations, which in some winter months can melt out.

The day of snowmelt can also be projected (Figure 37), and achieved by imposing a temperature change on the basin and its snowpack by shifting the corrected SCA fractions (Figure 16) up by 300 m per 2°C change in temperature, which is consistent with the current observations that each 300 m corresponds to a 2°C shift in temperature (Figure 11). A 2°C shift in temperature also results in a 300-m upward shift in the degree days (Figure 33). The effect is that with an increase in temperature, there is less snow and it melts faster. The magnitude of change in a given year depends on the patterns of SCA and SCA depletion that were observed. Effects



**Figure 35.** Cumulative degree day for current conditions, based on Figure 28; and cumulative degree day with temperature increases of 2-6°C. Lines represent means and shaded areas standard deviations based on inter-annual variability.



**Figure 36.** Winter SCA by elevation, current and projected with increased temperature. Error bars indicate inter-annual variability for nine years.

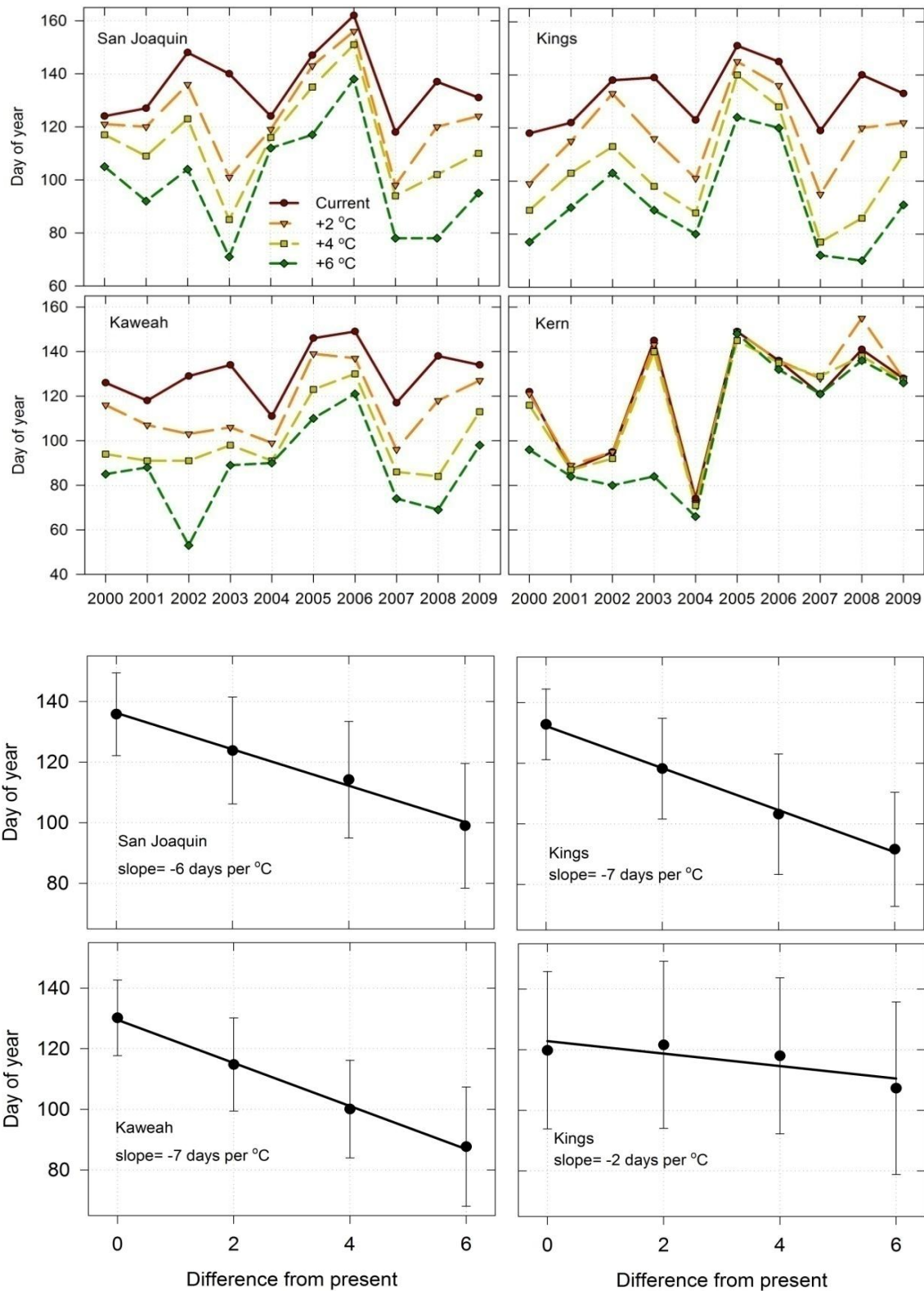
are similar for the 3 basins (San Joaquin, Kings, and Kaweah) (Figure 37). Overall, the timing of snowmelt shifts toward earlier in the spring at a rate of -6 to -7 days per °C. That is, with a temperature increase of 2°C, 50% of the seasonal snowmelt will be depleted about 2 weeks earlier than present. Of interest is that the imposed 6°C increase in temperature in 2005 and 2006 (wet year) are similar to conditions that were experienced in 2004 and 2007 (dry years). For example, in 2004 and 2007, in the Kings River basin, 50% of the seasonal snowcover was depleted by day 120, and with an imposed temperature shift of 6°C in 2005 and 2007 the date at which the snowcover was 50% depleted, shifted from day 150 and 145, respectively to about day 120.

Areas with persistent snowcover on May 1 or June 1 for 11-12 years (Figure 29c, 29d) are the areas with least potential for change. These are the areas with the most accumulation and/or slowest melt rates. Areas with persistent snowcover for 6 or fewer years on May or June 1 would be more vulnerable to change; and with a warming climate would be expected to have partial snowcover earlier in the winter and spring as well. Areas with snowcover for 6 or fewer years on April 1 are the most vulnerable to change. Note that the threshold used here is 20%, though the actual snow-covered can be expected to be higher in forested areas, owing to the limited ability of satellites to detect snow under tree canopies.

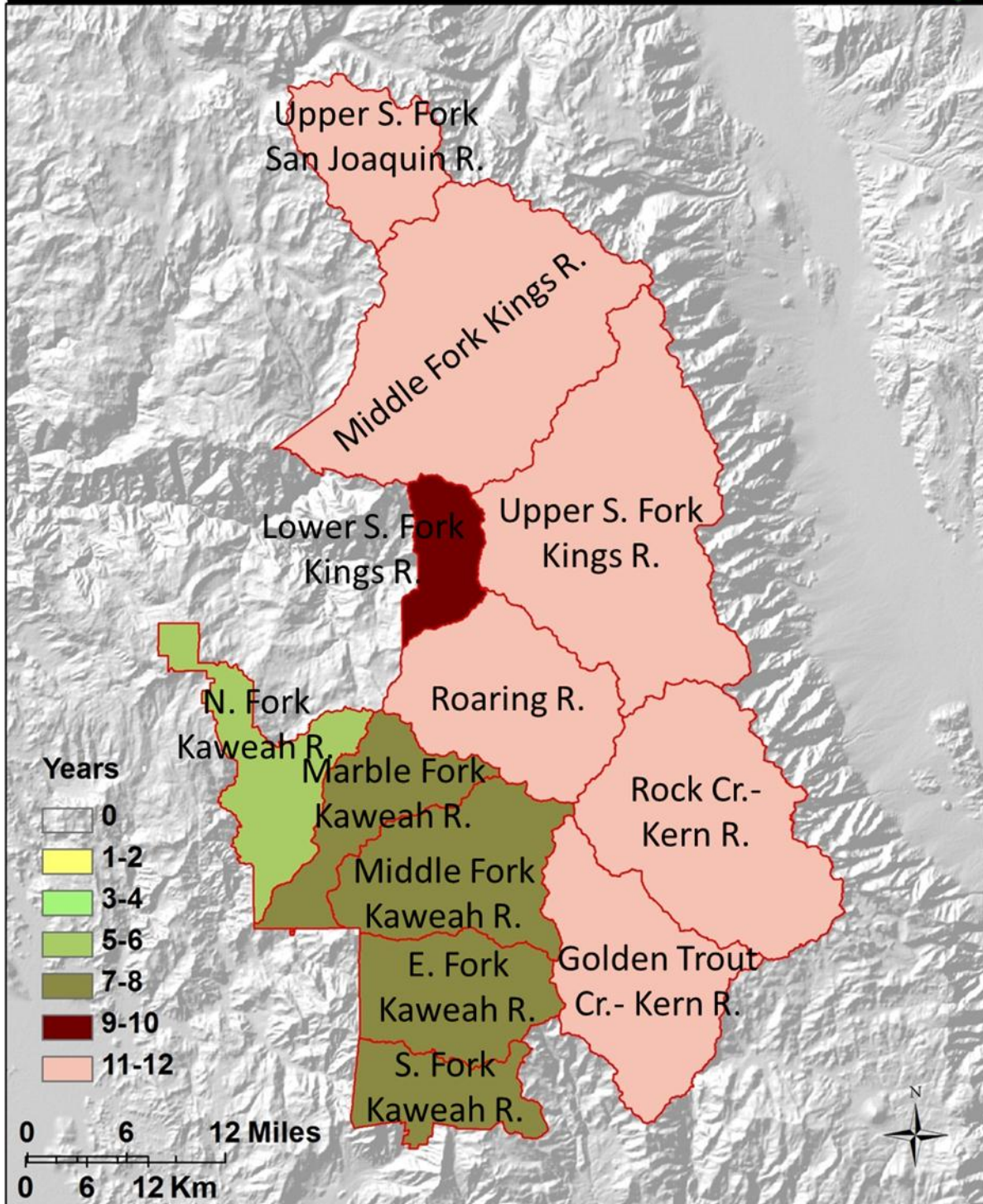
The southwestern HUC-10 watersheds that overlay the parks are the most vulnerable to lower snowcover and earlier snowmelt with climate warming, as they currently have persistent snowcover in March-April only about half to two thirds of the time (Figure 38a, 38b), dropping to lower than one-third of the time in some areas for May-June (Figure 38c, 38d). Note that these persistence maps are based on a 20% threshold for snowcover, which is just above the lower limit that can be detected with confidence from satellite. There is also considerable variability within each HUC-10 polygon, as elevation, aspect and forest cover, the main variables controlling snow persistence, vary tremendously within each.

Elevation patterns of snow persistence are apparent looking across elevation bands in all HUC-10 polygons (Figure 39). Note that in March areas above 2100 m generally met the 20% snowcover threshold used in this analysis in most years. That elevation increases to 2400 m in April, 2700 m in May and 3600 m in June. The elevations with these same thresholds can be expected to go up about 300 m for each 2-3°C increase in average temperature.



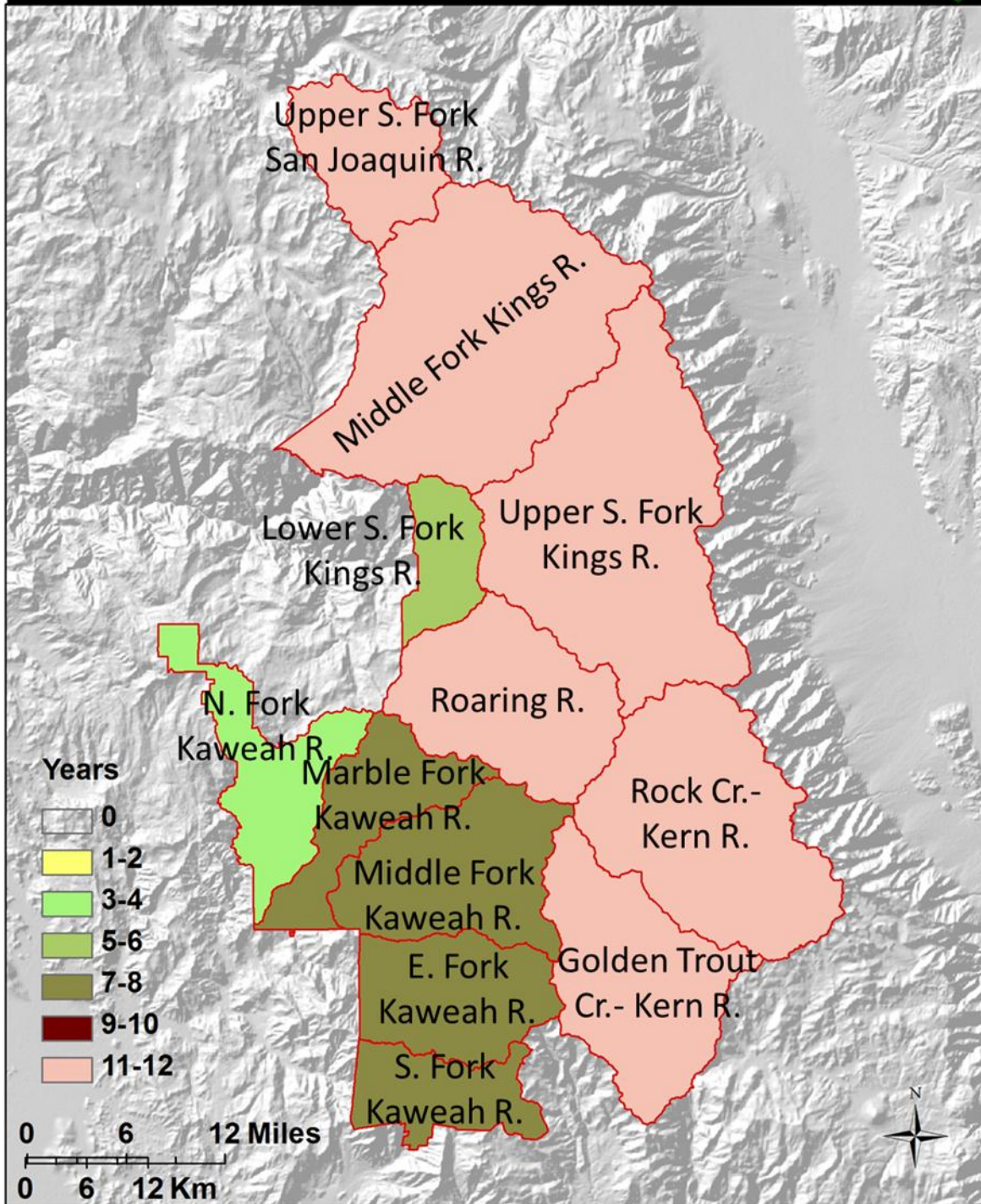


**Figure 37.** Snowmelt timing based on depletion method for current conditions, and with a 2, 4 and 6°C increase in average temperature. Upper panels shows day of year that 50% of seasonal snowmelt occurred; lower panels shows mean, best fit and standard deviation of the nine years.

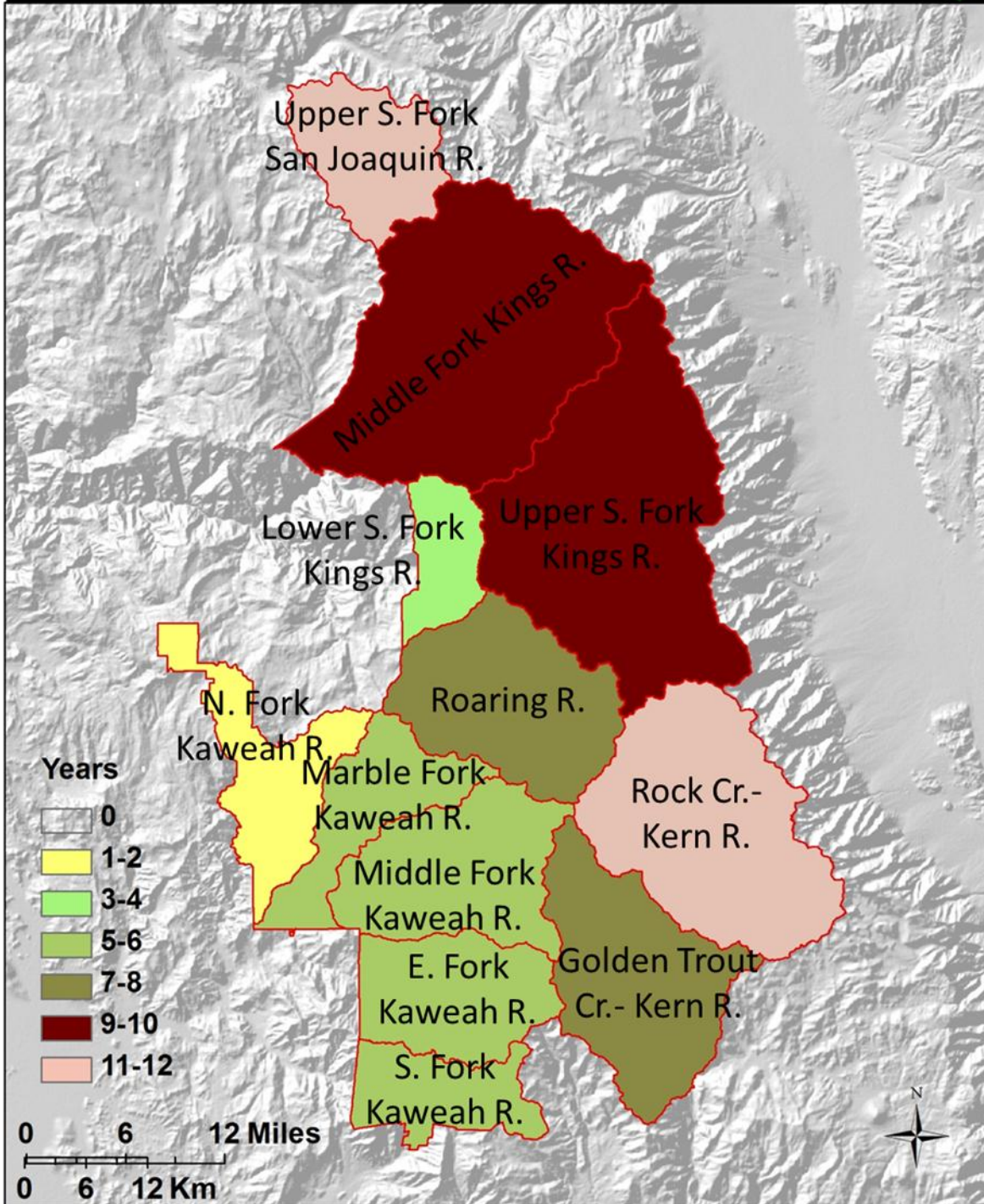


**Figure 38a.** Average snow persistence for HUC-10 polygons, March. Data based on Figure 29a. Snowcover persistence, or number of years each polygon had >20% snowcover on March 1.



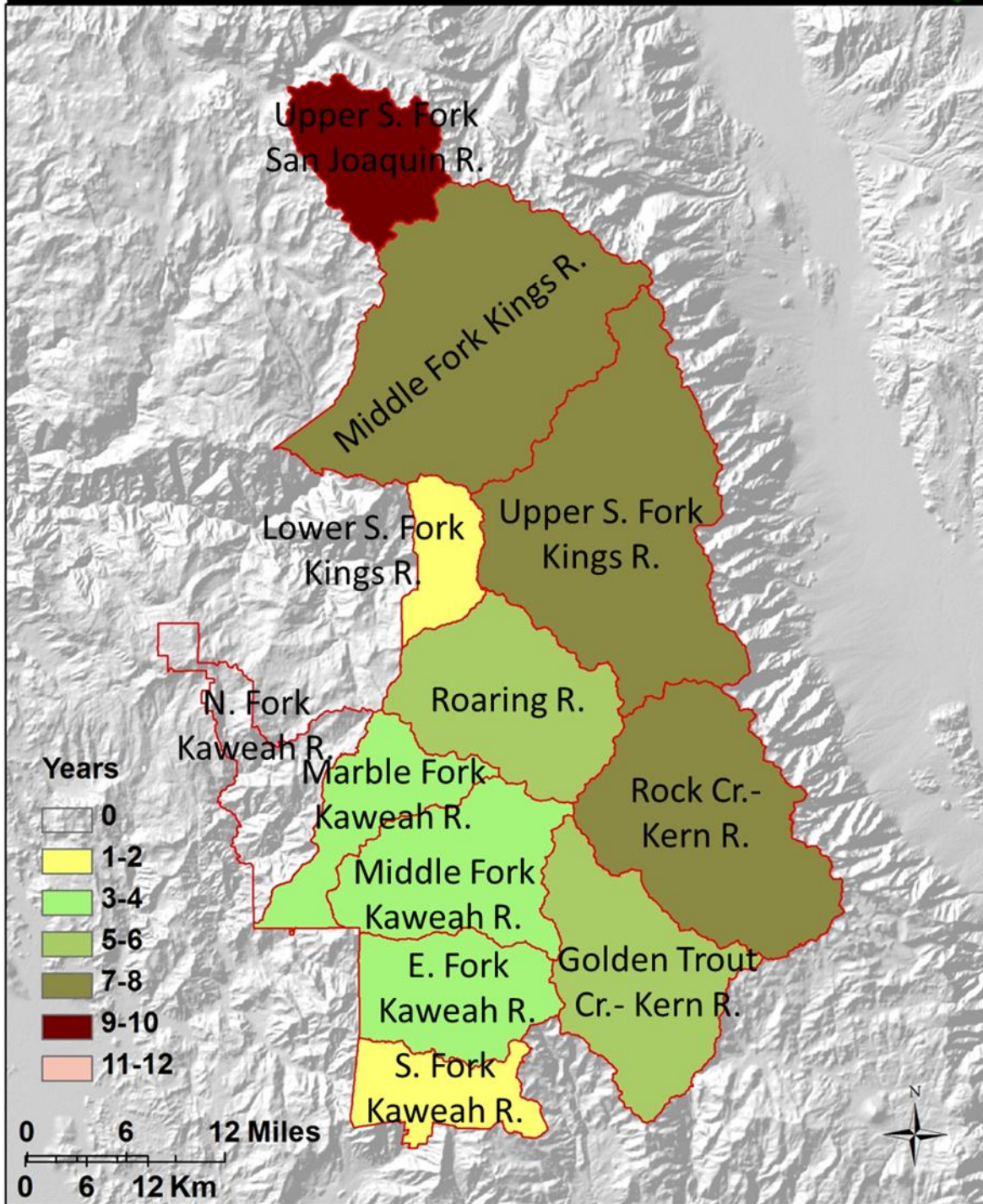


**Figure 38b.** Average snow persistence for HUC-10 polygons, April. Data based on Figure 29b. Snowcover persistence, or number of years each polygon had >20% snowcover on April 1.

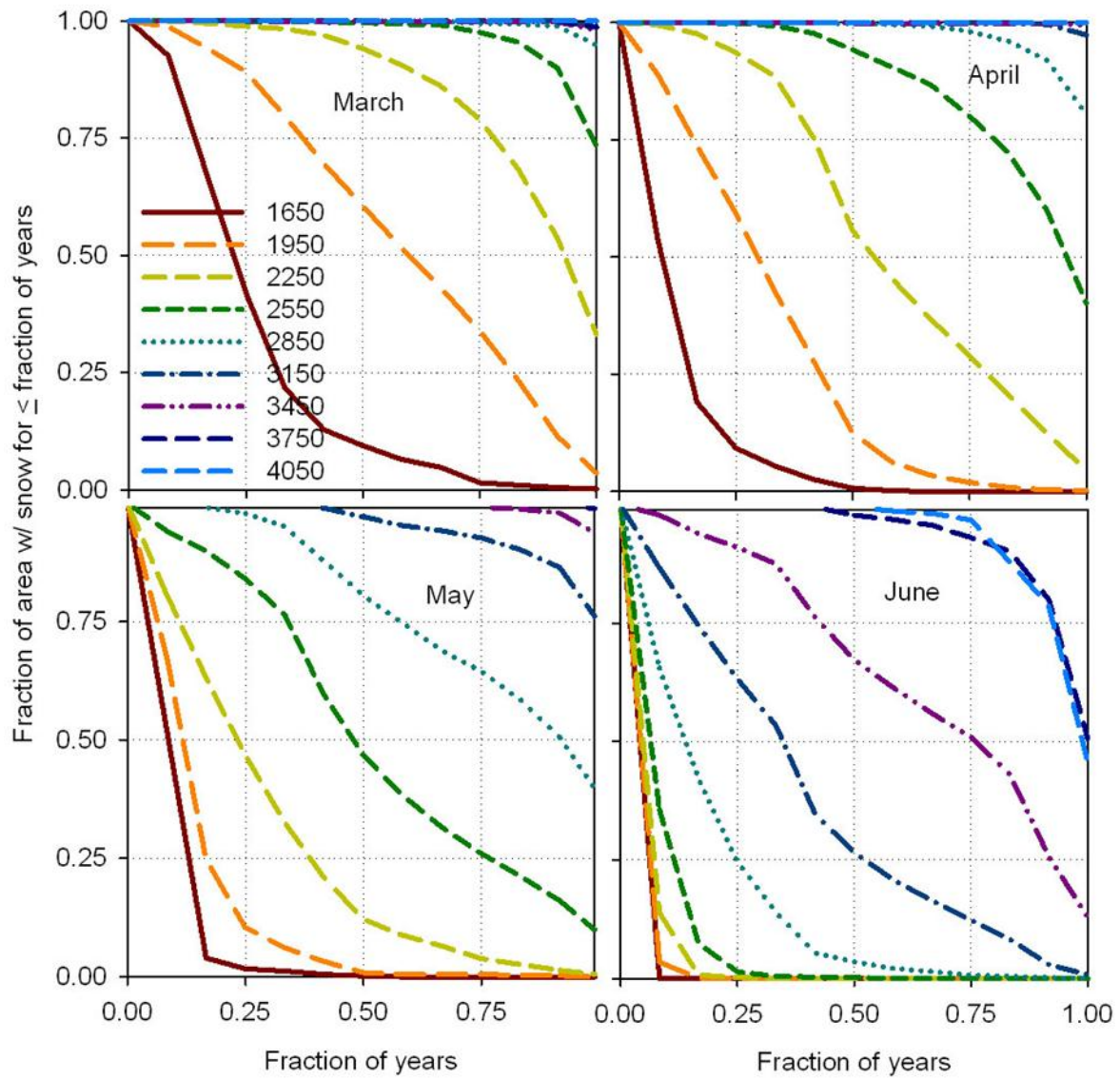


**Figure 38c.** Average snow persistence for HUC-10 polygons, May. Data based on Figure 29c. Snowcover persistence, or number of years each polygon had >20% snowcover on May 1.





**Figure 38d.** Average snow persistence for HUC-10 polygons, June. Data based on Figure 29d. Snowcover persistence, or number of years each polygon had >20% snowcover on June 1.



**Figure 39.** Snow persistence by elevation. Data based on same data as Figures 29a-29d. Abscissa refers to fraction of the 12 years of record and ordinate to fraction of area within that 300-m elevation band that had  $> 20\%$  snowcover. Center of each elevation band given in legend (meters)

## **Level of confidence in assessment**

Temperature projections have a relatively high level of confidence at the broad landscape scale, for both current and projected conditions. This is in part because elevation is a main determinant of temperature. However, the available data do not capture more-local variability, and current modeling is much too coarse to provide accurate estimates of the effects of local topography, e.g. valleys and ridges. Thus while available temperature data do indicate the general snowpack accumulation and melt patterns, they are not at a level of detail that permits predicting or downscaling snow patterns at the scale of tens to hundreds of meters. Thus broad-scale temperatures, averaged over elevation zones, have a good level of confidence; but local variability in temperature due to the complex topography has a much higher level of uncertainty.

Precipitation has a lower confidence for both the broad-landscape and more-local scales, as evidenced by the different patterns shown for PRISM estimates, station data and snowmelt estimates. Also missing is a credible estimate of rain versus snow fall across the study area, for daily, monthly, and annual times. Overall, the level of confidence for precipitation should be judged to be low, with PRISM estimates giving only broad-scale indications of precipitation, and station data having local relevance. The greatest potential for developing more-confident values from existing data is in the zone dominated by snow

## **Gaps in understanding**

SEKI has a fundamental gap in data and information about the water cycle, and how it affects water-dependent ecosystem services both in and downstream from the Parks. Accurate knowledge of fine-scale temperature, precipitation, and snowpack patterns are fundamental to understanding water- and temperature-dependent resources. It should be concluded that a current major gap in any resource assessment is the lack of understanding of the amount of precipitation occurring across the landscape, and the partitioning of this precipitation between rain and snow. Other components of the hydrologic cycle (Figure 2) have similar gaps. While improvements in modeling or gridding available data could help address these gaps, additional data for evaluation are also a clear need. Finer-scale analyses of paleoclimate data are also lacking. While we did not go back to the source data for the paleoclimate trends, an initial review of sources suggests that there is also a gap in spatially representative paleoclimate data for most areas within SEKI.

## **Recommendations for future study/research**

SEKI has a fundamental need for a modern program of measurement of water-cycle attributes and analysis of water-cycle data to support management of other resources and assessment of ecosystem services provided within and outside its boundaries. Developing an understanding of water-cycle resources specific to the relatively small sub-basins and complex topography of the SEKI watershed and should consider the following elements:

1. An expanded and enhanced measurement program of mountain temperature, precipitation, snowpack, soil moisture and selected other energy-balance components to

augment the current modest system, which was put in place over the past several decades and is largely based on technology developed over 50 years ago. Technological advances of the past decade have resulted in many low-power, low-cost, robust sensors and networks that can provide real-time data to inform operations, as well as longer-term information to inform planning and policies. At least one elevation transect of sensors, distributed over the landscape every 300 m, to capture variability, should be placed in each of the four distinct river basins in SEKI. Given the potential value of this information to downstream water users and to researchers, as well as to Park resource managers, opportunities for partnerships should be explored. A program such as is envisioned would involve hundreds of sensors distributed across the main snow- and runoff-producing sub-basins should have a very short payback in terms of the value of the information. Design of this network should consider vegetation, aspect, and radiation differences as well as elevation in placing sensors.

2. Improved characterization of forest vegetation from satellite and aircraft data in support of hydrologic data analysis and modeling. This should include complete LIDAR and hyperspectral coverage of forests sufficient to recover forest attributes such as density, tree height, leaf area index, and canopy structure.
3. Improved characterization of soils in SEKI, including depth and physical properties. Development of this information should be a major step in reducing uncertainty in water availability for ecosystems, and drought stress. Traditional physical approaches for soil characterization should be augmented with seismic or other geophysical surveys to provide spatial information.
4. Improved paleoclimate data and analysis. Existing data should be combined with locally relevant climate data from snow, temperature and streamflow measurements. There exist excellent opportunities to reconstruct multiple records specific to the basins in the Parks from the wealth of tree-ring and other paleo-climate records available in the basin, as has been done for multiple locations across the West (e.g. Meko and Woodhouse, 2005). This reconstruction is best done using the original tree-ring and other paleoclimate measurements, directly with the independent variable of interest, e.g. precipitation or streamflow basins. However, because this streamflow is correlated with other reconstructed records, some scenarios of wet periods can be developed from these other reconstructions. The exact experimental design for further measurements should follow an assessment of currently available data. Because these data will also be highly relevant to downstream water managers, opportunities for partnerships should be explored, e.g. with the Integrated Regional Water Management groups. Although modeling regional climate change reliably remains a challenge, it should be noted that the explanations available so far for decadal to century-scale hydroclimatic variability in recent centuries are similar to the mechanisms proposed for future regional change. Thus, the further analysis of the paleoclimate record will provide much-needed scenarios for how the hydro-climatological and ecological systems of the Sierra Nevada watersheds will experience and respond to climate change.
5. Further development of a more-accurate snowcover product. Although beyond the scope of the current analysis, SEKI actually has the energy-balance data needed to carry out a more-rigorous reconstruction of daily snow water equivalent, and hence precipitation and snowmelt for the four river basins within its boundaries. This product will then serve as



the most accurate spatial indicator of water resources currently available anywhere in the Sierra Nevada, and will serve as the foundation for modeling of the other components of the water balance, e.g. ET and soil moisture.

6. A rigorous, focused analysis of temperature and precipitation, and regridding to produce a PRISM-like data set would provide greater confidence. Additional quality control of data, plus a focused regional analysis is recommended. The PRISM algorithm or similar approach can be adapted to the snow, precipitation and temperature data of the SEKI region, following further detailed cleaning of data.
7. An ongoing program of assessment of hydrologic conditions, trends, forecasts, and outlooks, with particular emphasis on extreme events and years; integrating this program with decision support. That program will require an efficient data and information system for SEKI and related data. Data should be evaluated at least annually, and thorough assessments done every 3-5 years.

Planning infrastructure and operations for climate change needs to be a process and not just an event (Bales et al., 2004), and should involve a continuing program of measurement, modeling and assessment. Going forward, management of water-dependent resources should consider an adaptive-management approach, involving a continual cycle of investigation and synthesis to inform decision making.



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